

CRETACEOUS MAFIC DYKE SWARM, PEARY LAND, NORTHERNMOST GREENLAND: GEOCHRONOLOGY AND PETROLOGY

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ABSTRACT

A group of dyke rocks from the Frederick E. Hyde Fjord area form the southern termination of a prominent dyke-swarm in Peary Land, northernmost Greenland. The melanocratic dykes are NS- to NW–SE-trending, vertically dipping, ≤ 10 –12 m wide, and have narrow chilled margins. Whole-rock $^{40}\text{Ar}/^{39}\text{Ar}$ dating of three dyke rocks indicates emplacement at ca. 85 Ma, some 15 to 20 Ma older than previous K–Ar whole-rock dating indicated. Whole-rock $^{40}\text{Ar}/^{39}\text{Ar}$ dating of three Lower Paleozoic metasedimentary rocks collected hundreds of meters away from the dykes reflects partial resetting during dyke-rock emplacement and suggests that the area experienced a greater thermal disturbance than the outcrop distribution of the dyke rocks would suggest. The dyke rocks contain normally zoned, euhedral microphenocrysts of titaniferous (≤ 4 wt.% TiO_2) clinopyroxene in a matrix of olivine (Fo_{69-35}), plagioclase (An_{69-35} , ≤ 1.0 wt.% K_2O), clinopyroxene and ilmenite; rare matrix biotite [$\text{Fe}/(\text{Fe} + \text{Mg}) = 0.65$] occurs, and intergranular granophyre is present in the most evolved dykes. Whole-rock geochemistry indicates a basaltic composition, with the least evolved rocks containing 5.7% MgO , 14% FeO_T , 4% TiO_2 , 2.8% Na_2O , and 0.9% K_2O (by weight). In terms of major elements, the rocks correspond to sodic-type alkali basalts; the normative mineralogy indicates an affinity to both olivine tholeiites and alkali basalts. Trace-element contents (45 ppm Ni, 45 ppm Co, 80 ppm Cr, 335 ppm V, on average) and mg numbers between 0.42 and 0.45 indicate that the rocks do not represent primitive magmas. Abundances of the rare-earth elements ($\Sigma\text{REE} = 170$ to 200 ppm, $\text{La}_N = 100$) and strongly fractionated patterns ($\text{La}_N/\text{Lu}_N = 10$) are consistent with the alkaline nature of the dyke rocks. The lack of negative Eu anomalies in chondrite-normalized REE profiles is consistent with the absence of plagioclase phenocrysts, suppressed by a high H_2O content of the magma. Normalized extended trace-element diagrams and $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic plots indicate affinities with OIB. Although the $^{87}\text{Sr}/^{86}\text{Sr}_i$ values (0.70467 ± 10) do not indicate crustal contamination, $\delta^{18}\text{O}$ values (+6.0 to +7.9‰) are above typical mantle values and suggest some crustal influence.

Keywords: lamprophyres, dyke rocks, Cretaceous, Greenland, mantle plume.

SOMMAIRE

Un essaim de filons dans la région du fjord Frederick E. Hyde constitue la terminaison vers le sud d'un essaim important de filons à la Terre de Peary, dans l'extrême nord du Groënland. Les filons mélanocratiques sont orientés NS à NO–SE et verticaux, d'une largeur jusqu'à 10 ou 12 m, et possèdent une étroite bordure figée. Une datation $^{40}\text{Ar}/^{39}\text{Ar}$ de trois échantillons (roches totales) indique une mise en place à environ 85 Ma, c'est-à-dire quelques 15 ou 20 million d'années avant l'âge de mise en place préconisé d'après les datations antérieures par K–Ar sur roches totales. Une datation $^{40}\text{Ar}/^{39}\text{Ar}$ de trois échantillons de roche métasédimentaire d'âge paléozoïque inférieur, prélevés à des centaines de mètres des filons, indique un ré-équilibrage partiel au cours de la mise en place des filons. La région aurait donc subi un événement thermique plus répandu que ne l'indique la distribution des filons. Les roches ignées contiennent des microphénocristaux idiomorphes normalement zonés de clinopyroxène titanifère ($\leq 4\%$ TiO_2 , poids) dans une pâte contenant olivine (Fo_{69-35}), plagioclase (An_{69-35} , $\leq 1.0\%$ K_2O , poids), clinopyroxène

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et ilménite; la biotite $[\text{Fe}/(\text{Fe} + \text{Mg}) = 0.65]$ est plutôt rare dans la pâte, et une intercroissance granophyrique est présente dans les filons les plus évolués. La composition globale des roches est basaltique, les roches les plus primitives contenant 5.7% MgO, 14% FeO_T , 4% TiO_2 , 2.8% Na_2O , et 0.9% K_2O (poids). En termes des éléments majeurs, les roches correspondent à des basaltes alcalins à tendance plutôt sodique. La minéralogie normative laisse entrevoir une affiliation à la fois aux basaltes tholéïtiques à olivine et aux basaltes alcalins. D'après les teneurs en éléments traces (45 ppm Ni, 45 ppm Co, 80 ppm Cr, 335 ppm V, en moyenne) et les valeurs de l'indicateur mg, entre 0.42 et 0.45, les roches ne représenteraient pas des magmas primitifs. Les concentrations de terres rares (entre 170 et 200 ppm, $\text{La}_N = 100$) et les spectres fortement fractionnés ($\text{La}_N/\text{Lu}_N = 10$) concordent avec la nature alcaline des roches filonniennes. D'après l'absence d'une anomalie négative en Eu dans les profils de terres rares normalisés par rapport aux chondrites, le plagioclase n'aurait pas formé des phénocristaux, à cause de la teneur élevée en H_2O du magma. Les profils normalisés des éléments traces et les relations entre les rapports isotopiques $^{87}\text{Sr}/^{86}\text{Sr}$ et $^{143}\text{Nd}/^{144}\text{Nd}$ indiquent une affinité avec les basaltes des îles océaniques. Quoique les valeurs du rapport initial $^{87}\text{Sr}/^{86}\text{Sr}_i$ (0.70467 ± 10) ne témoignent pas d'une contamination par la croûte, les valeurs de $\delta^{18}\text{O}$ (+6.0 to +7.9‰) surpassent les valeurs typiques du manteau, et font penser que la croûte a exercé une certaine influence.

(Traduit par la Rédaction)

Mots-clés: lamprophyres, roches filonniennes, âge crétacé, Groënland, panache du manteau.

INTRODUCTION

The northern part of Greenland is underlain by a thick sequence of folded Paleozoic sedimentary units that form the eastern continuation of the Franklinian Basin of the Canadian Arctic (Dawes 1971). Superimposed on the Paleozoic rocks are Cretaceous–Tertiary events, including a swarm of diabase dykes and a sequence of peralkaline volcanic rocks (the Kap Washington suite), that reflect plate movements and dynamics occurring in conjunction with formation of the surrounding ocean basins (Soper *et al.* 1982, Batten *et al.* 1981). A study of the Cretaceous dyke swarm is relevant with respect to the interpretation of basin development of northern Greenland, but the age, petrogenesis and tectonic setting of the dyke rocks remain ambiguous, as does their timing relationship to the peralkaline volcanism. In this paper, results of high-precision geochronology ($^{40}\text{Ar}/^{39}\text{Ar}$ whole rock) and the first isotopic (Nd, Sr, O) data obtained on the dyke rocks are presented, along with additional information on whole-rock and mineral chemistry, in order to define more precisely their petrogenesis. The results suggest that the dyke rocks are older than previously considered, and have chemical affinities to ocean-island basalts (OIB). These data are cautiously interpreted to reflect the former presence of a mantle plume, thus supporting recent suggestions that a large igneous province once occupied the high Arctic area (HALIP) and that it formed above a long-lived Cretaceous mantle plume (Tarduno 1998, Maher 2001). If this inference is correct, the results have important implications with respect to the tectonic evolution of northern Greenland during the Cretaceous.

We emphasize here that the present study resulted from a regional reconnaissance program designed to assess the mineral potential of this very remote and inaccessible study-area, in particular potential for Zn–Pb SEDEX mineralization given the presence of the large Citronen Fjord Zn–Pb deposit (van der Stijl & Mosher

1998). During the course of this reconnaissance work, outcrops of mafic dyke rocks were sampled for subsequent study. Our intent in this paper is to report on these results in light of their significance to the origin of a much more extensive occurrence of contiguous dyke-rock to the north of the Frederick E. Hyde Fjord area (Fig. 1).

PREVIOUS WORK AND SAMPLING

General reference has been made to mafic dyke-rocks in northernmost Greenland as part of regional mapping programs. These dykes post-date the youngest structural event recorded in the host rocks, either folded Paleozoic sedimentary units or post-orogenic Permian–Carboniferous sediments (*e.g.*, Dawes & Soper 1970, Dawes 1971, Pedersen 1980, Friderichsen & Bengaard 1985). The most thorough account of the distribution of the dykes was given by Higgins *et al.* (1981), who noted that they are generally N–S-trending and increase in density west of Polkorridoren (Fig. 1). In some areas, the dykes account for 20% of outcrop and, whereas most are 3 to 30 m wide, a number are 200 m wide. There is a change in the trend of the dykes toward an E–W direction southward across Amundsen Land, which coincides with a dramatic decrease in the frequency of dykes (Fig. 1), as is typical of dyke swarms in general (Ernst *et al.* 1995). Higgins *et al.* (1981) were not sure if the change in orientation of the dykes was a primary or secondary feature. These authors inferred a general Cretaceous–Tertiary age for the dykes and noted their importance with respect to evolution of the adjacent ocean basins. Previous petrological work on the dykes documented them as ophitic diabase and olivine diabase with an alkaline character (Soper *et al.* 1982).

The first reported age for a dyke rock from the study area was a conventional K–Ar age of 72.2 ± 9.0 Ma (average of three analyses) for a NW–SE-trending olivine diabase dyke collected south of the Midsommersø

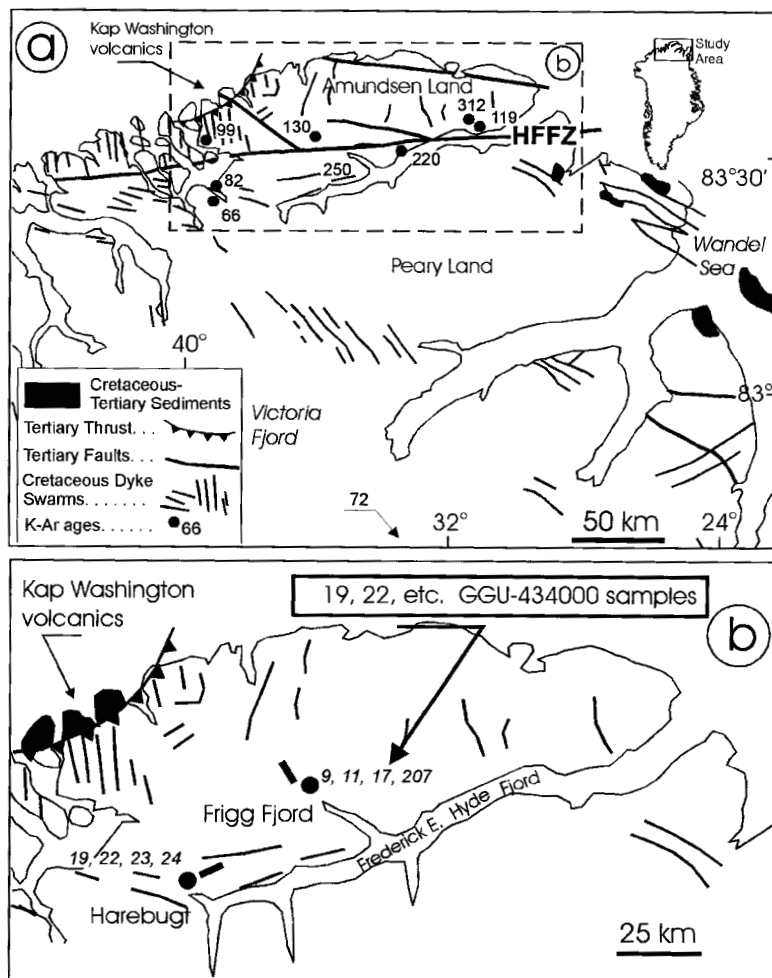


FIG. 1. (a) Location of the study area in northernmost Greenland, distribution of mafic dykes, considered to have formed during a Late Cretaceous – Early Tertiary magmatic event, and ages of dyke rocks previously dated by conventional K–Ar method (see text). (b) Inset map from Figure 1a showing locations of dyke rocks collected in this study. The large bars indicate the general trend of dykes, and the numbers (prefix GGU–) refer to the samples discussed in the text. HFFZ is the Harder Fjord Fault Zone.

area ($82^{\circ}12' \text{ N}$, $33^{\circ}03' \text{ W}$), southern Peary Land (Fig. 1; Bridgwater 1970, Henriksen & Jepsen 1970), whereas a second K–Ar age of $66.6 \pm 6.6 \text{ Ma}$ was obtained for a E–W-trending olivine diabase dyke on the south side O.B. Bøggild Fjord, southern Peary Land (Bridgwater 1971, Dawes & Soper 1971). Dawes *et al.* (1983) presented results of conventional K–Ar dating of nine dyke rocks, including the previously mentioned 66 Ma sample (Fig. 1). The dated samples range from 511 Ma to 66 Ma, and there is no apparent trend to the ages. Given the unrealistically old age for one sample and the

large range in ages for the others, Dawes *et al.* (1983) concluded that there had been disturbance of the K–Ar system and that excess argon was probably an inherent problem. The authors considered the younger ages of 82, 72 and 66 Ma to be the most reliable, given that they come from a relatively less deformed part of the fold belt.

Recently, on the basis of paleomagnetism, Abrahamsen *et al.* (1997) subdivided the dykes into two age groups, namely (1) Carboniferous dykes corresponding to the most northerly occurrences, with a

dominantly N–S orientation, and (2) Cretaceous (57 ± 10 Ma) dykes that have NW–SE and E–W orientations. The dykes do not seem to cross-cut the Late Cretaceous – Early Tertiary (*i.e.*, Maastrichtian–Danian) Kap Washington volcanic rocks (Brown & Parsons 1981, Larsen *et al.* 1978, Håkansson *et al.* 1991).

Dyke rocks north of Frederick E. Hyde Fjord off the northern ends of Frigg Fjord and Harebugt (Fig. 1) have been sampled. At Frigg Fjord, samples are from four localities (GGU–434009, 434011, 434017, 434207), with the dykes ranging from 1–2 to 10–15 m in width and oriented 135° to 155°N, dipping 50° to 90° SE. All of the dyke samples are fine grained except for GGU–434207, which has a distinct ophitic texture. Samples were collected from chilled margins apparently free of visible alteration, except for GGU–434207, which comes from the interior of a 15-m-wide dyke. Samples from Harebugt (GGU–434019, 434022, 434023, 434024) are similar to the previous dykes, but are oriented 060° to 080° N, with dips 60° to 90°S. Sample GGU–434019 is from a chilled margin and seems similar texturally to dykes from Frigg Fjord. Samples GGU–434022 (margin) and GGU–434023 (core) are from the same dyke, with the latter containing coarse euhedra of calcite infilling vugs (see below); these samples, along with GGU–434024, are distinctly coarser grained than the other dykes. At localities 434017 and 434207 the dykes are associated with local development of abundant epidote – carbonate – quartz – chlorite, either as marginal alteration or veinlets within the dykes. Along the margin (*i.e.*, several m) of all the dykes, the sedimentary rocks are highly indurated as a result of contact metamorphism.

PETROGRAPHY OF DYKE ROCKS

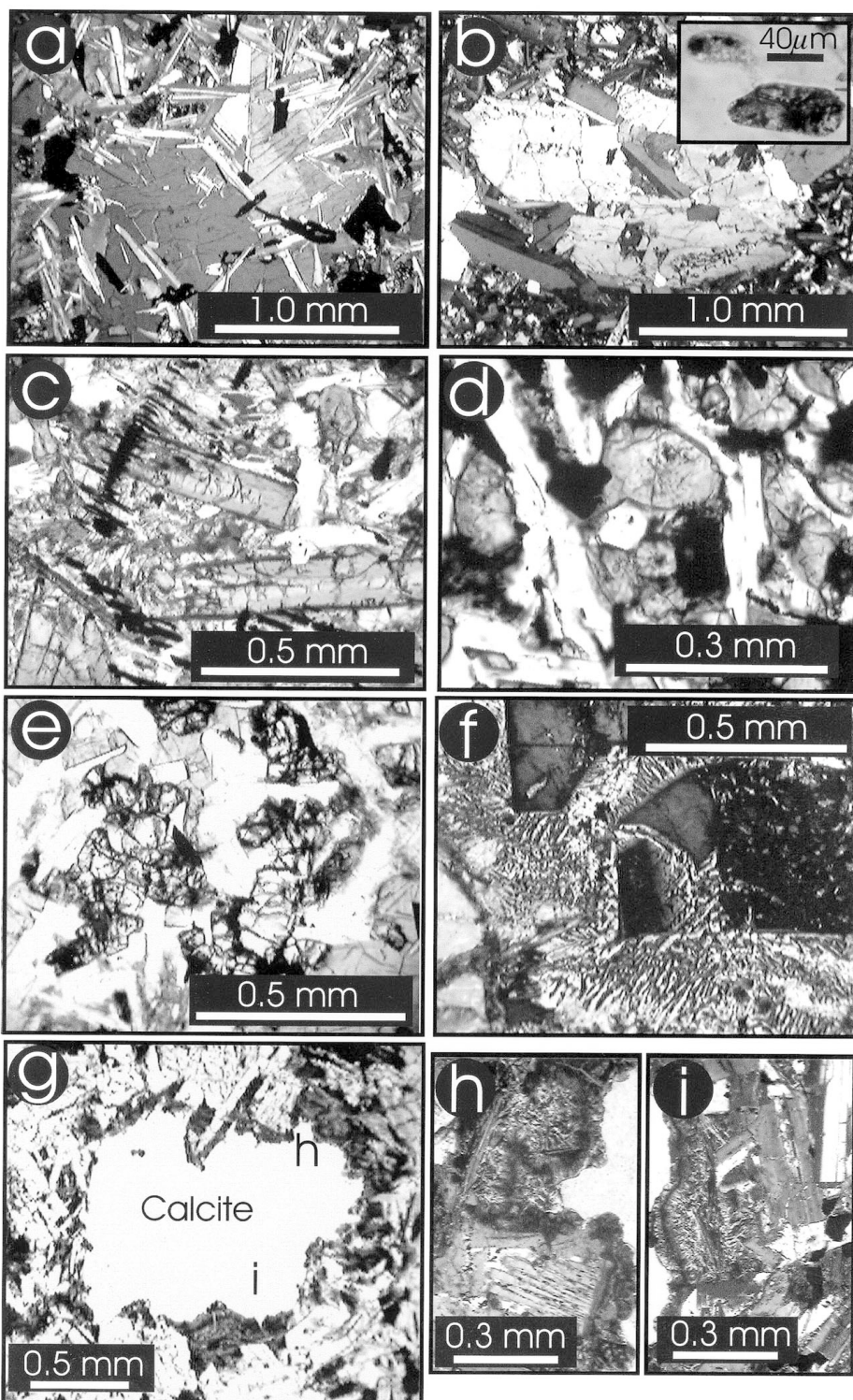
Petrographically, the dyke rocks are aphyric and vary from fine to medium grained, with the latter characterized by an ophitic texture (Fig. 2a). The following description pertains to all but one of the samples studied, described separately below. The dyke rocks are dominated by both medium- and fine-grained, euhedral to subhedral clinopyroxene (Figs. 2a, c, d) with both normal and oscillatory zonation and variably developed pink pleochroism. In some cases, primary melt inclusions decorate growth zones (Fig. 2b). Minor, anhedral olivine occurs in the matrix, where it is intergrown with clinopyroxene–plagioclase (Fig. 2e). Imaging analysis using the electron microprobe indicates a complex overgrowth of orthopyroxene and biotite around the olivine, and also veinlets of orthopyroxene cutting olivine (Fig. 3b). Olivine also occurs as remnant grains within a green-brown mass of secondary minerals (serpentine). Plagioclase (An_{60-70} to An_{40-50} composition by Michel-Lévy method) generally occurs as lath-shaped grains, but in coarser rocks it may be rectangular to equant (Fig. 2b). Rare occurrences of coarser plagioclase ± clinopyroxene clots, possibly glomeroclasts, occur in

medium-grained dyke-rocks. The clinopyroxene:plagioclase ratio does vary, but this may in part reflect flowage differentiation (*e.g.*, Dostal & Durning 1998). An oxide phase (*i.e.*, ilmenite) occurs as either equant, subhedral to euhedral grains or acicular to skeletal grains, and both types are confined to the matrix. Exsolution textures are absent within the oxides. Apatite is present in all dyke rocks as both fine-grained, acicular, radiating grains and as equant grains, and its abundance correlates positively with the amount of plagioclase. Biotite occurs as a subhedral to euhedral matrix phase in a single sample (GGU–434009), and texturally seems to be late primary and also replacing clinopyroxene.

The exception to the above petrographic description is sample GGU–434207, which is a granophyre-bearing plagioclase – amphibole – biotite ± quartz ± sanidine dyke-rock. The granophyre, which envelops strongly zoned plagioclase euhedra (Fig. 2f), contains coarse euhedra of apatite and subhedral to euhedral sanidine and albite. Rare phenocrysts of chloritized amphibole and coarse grains of chloritized biotite also are present in this dyke-rock.

Of particular note petrographically is the presence of spherical to ovoid cavities in a few dyke rocks (GGU–434023 and 434024). These cavities are occluded with carbonate and surrounded by secondary minerals after primary mafic phases (Figs. 2g, h, i). In addition, plagioclase laths are arranged concentrically about the cavities (Fig. 3a), suggesting a primary origin rather than a

FIG. 2. Photomicrographs of the dyke rocks from north Greenland. Photos a, b, f, h and i taken in crossed nicols. (a) GGU–434017: medium-grained, ophitic mafic dyke-rock with coarse subhedral grains of titaniferous augite intergrown with laths of calcic plagioclase and subhedral grains of ilmenite. (b) GGU–434024: glomeroclast-like aggregate of clinopyroxene and plagioclase in finer-grained matrix of clinopyroxene – plagioclase – oxides (Fig. 2c). Inset shows a melt inclusion that decorates the growth zones of clinopyroxene. (c) GGU–434024: fine-grained mafic dyke with euhedral laths of zoned clinopyroxene intergrown with euhedral plagioclase and late skeletal Fe–Ti oxides. (d) GGU–434011: Matrix of fine-grained mafic dyke with equant, euhedral grains of clinopyroxene intergrown with euhedral plagioclase and equant, euhedral Fe–Ti oxides. (e) GGU–434017: matrix to Figure 2a showing subhedral to anhedral grains of olivine surrounded by plagioclase and clinopyroxene. Note that the olivine is rimmed by orthopyroxene and biotite, as revealed by BSE imaging. (f) GGU–434207: zoned grains of calcic plagioclase surrounded by granophyric quartz + sodic plagioclase with minor additional quartz, sanidine and apatite. (g, h, i) GGU–434023: vesicle occluded by ferroan and manganoan calcite with altered mafic minerals rimming the vesicle and feldspar-rich matrix. See Figure 3a for a BSE image of this area.



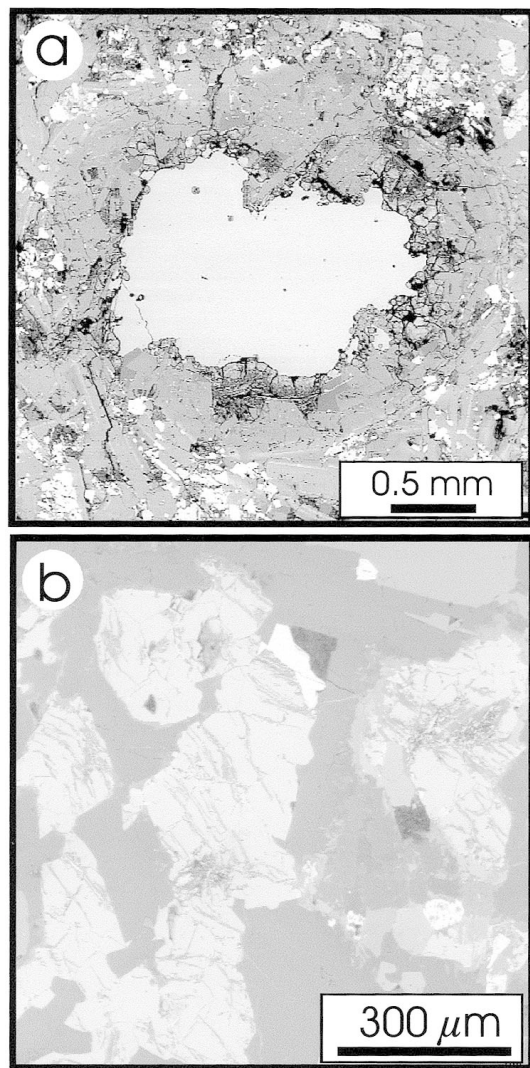


FIG. 3. (a) Back-scattered electron (BSE) image of mafic dyke-rock (GGU-434023) showing vesicle-type feature rimmed with secondary, hydrated Fe-Mg-Al silicates and occluded with Fe- and Mn-rich calcite. This feature may represent a primary fluid-escape feature relating to the volatile-rich nature of the magma. (b) BSE image of mafic dyke-rock (GGU-434017) showing subhedral grains of olivine (light grey) intergrown with plagioclase (dark grey). The olivine grains are cut by orthopyroxene-filled fractures.

later alteration-induced phenomenon. These features may represent gas-escape structures, as do the ocelli that occur in some alkaline rocks (*e.g.*, Mitchell *et al.* 1991, Le Roex & Lanyon 1998).

Alteration is variably developed in the rocks, with intensity increasing in rocks that are relatively more

feldspathic (*i.e.*, alkaline). Samples GGU-434009, 011, 017 and 019 have a relatively minor abundance of secondary phases (< 1% modally); samples GGU-434022, 023 and 024 have a few modal % alteration, and sample GGU-434207 is characterized by an abundance of secondary phases. For purposes of discussion these groupings are herein designated as groups 1, 2 and 3. We further note that group-1 rocks are also characterized by the presence of olivine. In general, secondary mineral phases include serpentine, carbonate, "leucoxene", "iddingsite", chlorite, biotite, and white mica. Anhedral to subhedral grains of pyrite commonly occur within areas of alteration. Further details of alteration are noted below under mineral chemistry.

ANALYTICAL TECHNIQUES

Powdered whole-rock samples were analyzed for major and trace elements (Cl, Sc, V, Cr, Co, Zr, Ni, Cu, Zn, Ba, Sr, Rb, Ga, Y, Nb, Sn, Pb) on fused and pressed pellets, respectively, using an automated Philips 2400 X-ray-fluorescence (XRF) spectrometer at St. Mary's University, Nova Scotia. In addition, concentrations of the remaining suite of trace elements (Li, Mo, Cs, Hf, Ta, Bi, Th, U) and rare-earth elements were established by solution chemistry using the ICP-MS facilities at Memorial University, St. John's, Newfoundland, following the procedures in Jenner *et al.* (1990). Analysis of the rocks for CO₂ (Leco carbon analyzer) and FeO/Fe₂O₃ (titration) were done at the Technical University of Nova Scotia, Halifax. Note that total H₂O was calculated as the difference between loss on ignition (LOI) and the CO₂ reported. Mineral compositions were determined on representative samples using an energy-dispersion system with a JEOL 733 Superprobe at Dalhousie University (Halifax, Nova Scotia) following procedures in Kontak (1995). The same facility was used to examine samples with back-scattered electron (BSE) imaging in order to detect chemical variation (*e.g.*, zoning).

⁴⁰Ar/³⁹Ar analyses of a whole-rock dyke and samples of metasedimentary rock were done at the Geochron laboratory of Queen's University (Kingston, Ontario) following the procedures outlined in Clark *et al.* (1998). Samples were crushed and 60 to 80 mesh fractions, sieved, washed and irradiated at the McMaster University nuclear reactor (Hamilton, Ontario) for analysis. In the case of the samples of dyke rock, several splits of the same sample were analyzed, including magnetic and nonmagnetic fractions, as described in the results below. Samples were step-heated using a defocused, 8W Lexel 3500 continuous argon ion laser. The evolved gas, after purification, was admitted to an on-line, Mass Analyzer Products 216 mass spectrometer for analysis. Errors for the individual steps, age spectra and isotope-correlation diagrams represent the analytical precision at a 2σ level, assuming the errors in the ages of the flux monitors are nil. This approach is suitable for a com-

parison of within-spectrum variation and a determination of which steps constitute a plateau (McDougall & Harrison 1988). A conservative estimate for the error in the J value is 0.5%.

Oxygen isotopic compositions of whole rocks and ^{18}O and ^{13}C for carbonates were determined at Queen's University using standard methods, as described in Kyser *et al.* (1998a, b), respectively. Radiogenic isotopes (Nd, Sr) were determined on two powdered samples at Memorial University of Newfoundland using procedures outlined in Jenner *et al.* (1990).

$^{40}\text{Ar}/^{39}\text{Ar}$ DATING

Whole-rock dyke sample GGU-434009, from Frigg Fjord, was analyzed in duplicate using the $^{40}\text{Ar}/^{39}\text{Ar}$ step-wise heating technique, on magnetic and non-magnetic fractions of the dyke (Table 1, Figs. 4a, b). [Table 1 is available from the Depository of Unpublished Data, CISTI, National Research Council, Ottawa, Ontario K1A 0S2.] The two $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra have U-shaped profiles, indicating the presence of excess argon; thus the integrated ages of ca. 160 Ma for these runs must be considered maximum ages. In the age spectra, the minimum ages correspond to the gas fractions with the lowest Ca/K ratios. In one run (Fig. 4a), a plateau age of 90.6 ± 4.0 Ma is defined, which is slightly younger than the minimum of 97 Ma for the other age spectrum (Fig. 4b). An isotope-correlation age of 84.7 ± 4.0 Ma (MSWD = 1.9, 55% of gas) is obtained for the non-magnetic fraction, but the magnetic fraction did not give a statistically valid correlation-age. A combined correlation-age for the two runs is 85.7 ± 4.7 Ma (MSWD = 4.9, 64% of gas; Fig. 4c). The best estimate for the age of the dyke is considered to be ca. 85 Ma.

A second sample (GGU-434013) from Frigg Fjord (Fig. 4d) was taken from the margin of an altered dyke with abundant secondary silica, epidote, carbonate and pyrite in the surrounding rock. The age spectrum, although somewhat irregular, indicates ages for most steps of ca. 75 to 86 Ma. The integrated age for this sample is 77.7 ± 0.7 Ma.

A third dyke rock (GGU-434019), from Harebugt, has a uniformly flat age-spectrum indicative of a simple thermal history (Fig. 4e). The plateau and integrated ages are identical at 81.9 ± 1.0 Ma. This age is similar to the sample from Frigg Fjord.

Whole-rock samples of slate from the same area at Frigg Fjord as the dyke rock samples also were analyzed by the $^{40}\text{Ar}/^{39}\text{Ar}$ step-wise heating method. The results (Fig. 5) are discussed here given the relevance of the data (see below) to the present study. Slate sample GGU-434001 (Fig. 5a), located several hundred meters from the nearest dyke-rock, has no obvious features of thermal overprinting, but the age spectrum shows a gradient from a minimum age of ca. 145 Ma for the low-temperature gas fraction to a maximum of ca. 500 Ma for the high-temperature steps. The second sample of

slate (GGU-434002; Fig. 5b), also well removed from any apparent dyke-rocks, also shows a step-wise age gradient, but with a minimum age of 226 Ma and high-temperature steps near ca. 400 Ma. A similar $^{40}\text{Ar}/^{39}\text{Ar}$ age pattern occurs for the third sample (GGU-434016; Fig. 5c), but the lowest-temperature gas fraction has a much younger age of 80 Ma, which is close to the inferred age for the dyke rocks (Fig. 4). The high-temperature gas fractions also approach ca. 500 Ma, with a maximum age of 662 Ma for the highest-temperature step. This particular sample is of note because it is from a zone of intense alteration (white mica) in the metasediments, and its degree of degassing may reflect the abundance of this secondary mineral. The $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra obtained for the slates are generally interpreted to reflect a thermal overprinting event, as documented in areas of contact metamorphism (e.g., Berger 1975, McDougall & Harrison 1988). By analogy, the age spectra for the slates samples suggest that a ca. 80 Ma thermal event overprinted rocks that had previously been outgassed by a thermal event at ca. 500 Ma.

CHEMICAL COMPOSITION

Composition of the primary minerals

The primary minerals in the dykes includes olivine, clinopyroxene, plagioclase, orthopyroxene, biotite, Fe-Ti oxides and apatite. Olivine (Table 2) is normally zoned from Fo₆₅₋₆₉ to Fo₃₅₋₃₉ without any discontinuities, as indicated by BSE imaging.

The compositions of the coarser grains of clinopyroxene and of the finer matrix grains are similar (Table 3), corresponding to augite and rare diopside in the Ca-Mg-Fe pyroxene plot (Fig. 6). Clinopyroxene

TABLE 2. REPRESENTATIVE COMPOSITIONS OF OLIVINE, NORTH GREENLAND DYKE ROCKS

Sample	GGU 434011	GGU 434011	GGU 434011	GGU 434011	GGU 434011	GGU 434017
Point	4	5	6	23	24	6
Notes	rim	core	core	matrix	matrix	rim
SiO ₂ , wt. %	33.02	36.07	35.51	36.68	34.76	32.81
FeO	47.38	33.15	35.02	27.47	28.15	46.46
MnO	0.86	0.47	0.48	0.45	0.44	0.80
MgO	16.56	28.82	27.72	32.92	30.38	17.69
Total	99.01	100.01	100.19	98.57	95.72	99.11
Structural formulae (4 atoms of oxygen)						
Si <i>apfu</i>	1.000	0.990	0.980	1.000	0.980	0.980
Fe	1.200	0.760	0.810	0.620	0.670	1.160
Mn	0.020	0.010	0.010	0.010	0.010	0.020
Mg	0.740	1.180	1.150	1.330	1.280	0.790
mole % Fo	38.3	60.7	58.5	68.1	65.7	40.4

The structural formulae are expressed in atoms per formula unit, *apfu*.

shows both normal and reverse zonation, and multiple zones may occur within larger phenocrysts, as observed by BSE imaging. We note that there is a distinct difference in general for pyroxene in group-1 and group-2 dyke rocks in the Ca–Mg–Fe plot (Fig. 6). This differ-

ence is also manifest in binary element plots that show a distinct separation of the pyroxenes in terms of their proportions of Si, Al, Fe and Ti (Fig. 7). The pyroxene in group-2 dyke rock is relatively enriched in Ca, Al, Ti and Si. Collectively, the data define continuous trends

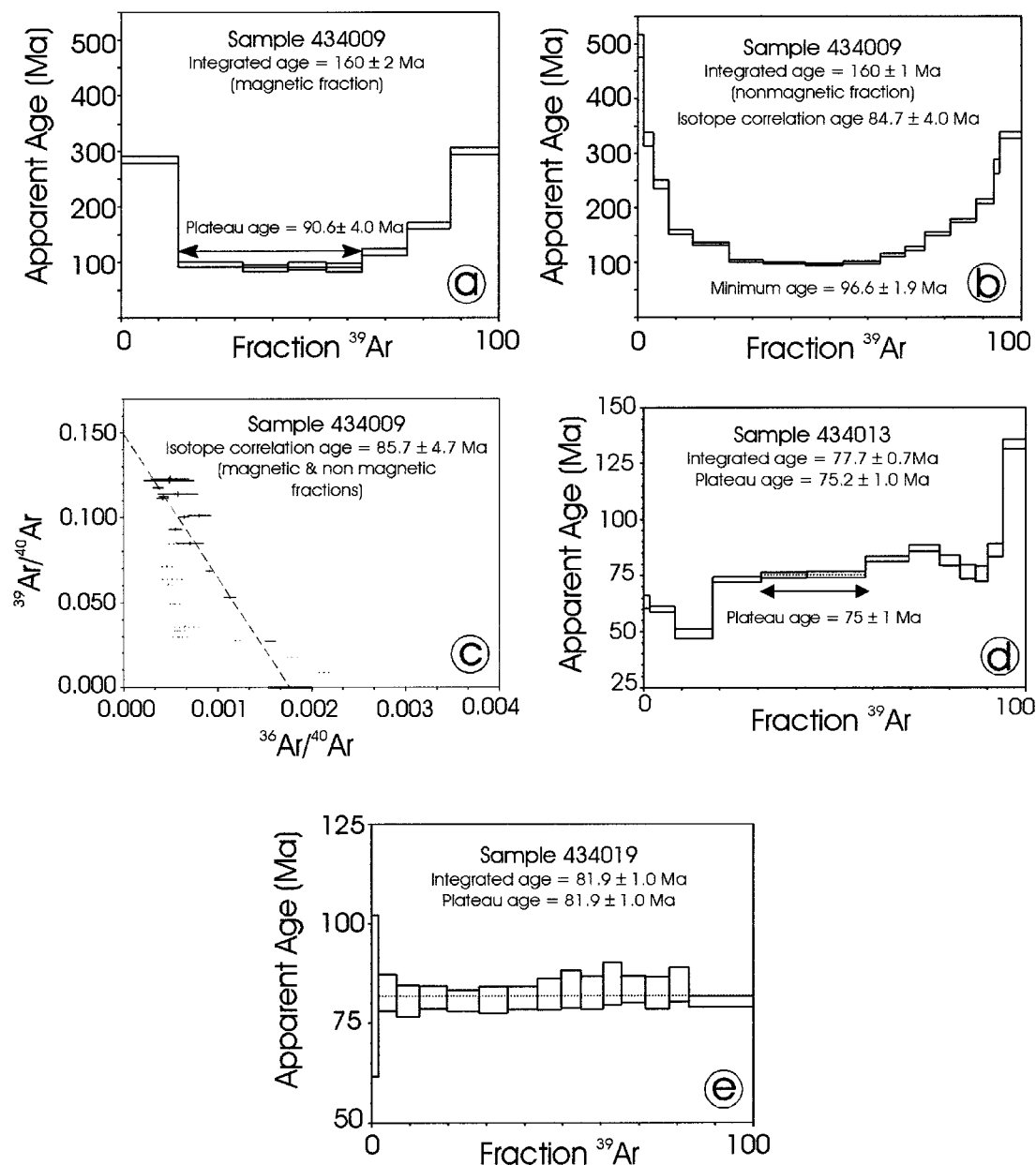


FIG. 4. $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra for dyke-rock samples GGU-43409 (a, b, c), GGU-434013 (d) and GGU-434019 (e): (a) magnetic fraction of dyke rock, (b) nonmagnetic fraction, (c) nonmagnetic and magnetic fractions in isotope-correlation plot, and (d, e) whole-rock samples.

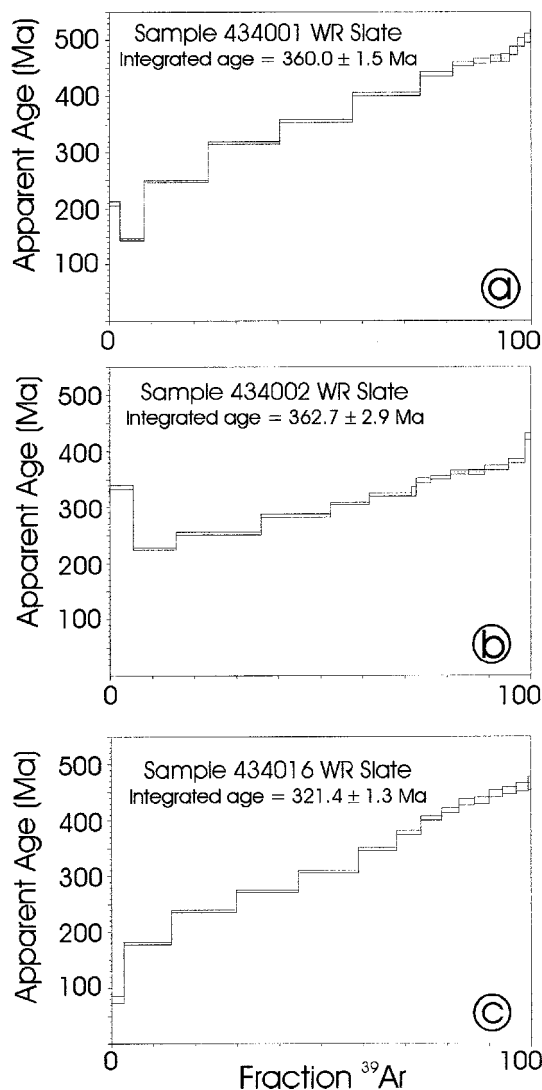


FIG. 5. $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra for metasedimentary rocks (i.e., slates). Note that each sample shows a monotonic gradient typical of samples that have experienced thermal overprinting. Note that sample GGU-434016 is characterized by intense alteration to white mica.

in the binary element plots, and there is an excellent correlation between Si and (Al + Ti + Ca) ($r = 0.98$; Fig. 7f).

Plagioclase phenocrysts and laths are compositionally similar and both are normally zoned, with core compositions An_{69} to An_{55} and rim compositions An_{35-55} (Fig. 8a, Table 4); no discontinuities have been noted

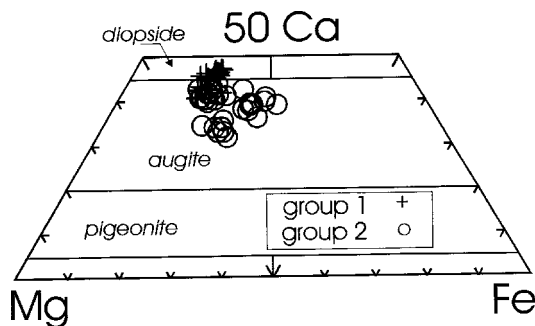


FIG. 6. Partial triangular plot (50Ca–Mg–Fe) showing compositions of clinopyroxene in mafic dyke-rocks using pyroxene nomenclature of Morimoto (1989). The groups refer to group-1 and -2 dyke rocks discussed in the text.

either from our petrographic study or by imaging with the electron microprobe. Plagioclase contains up to 1.6 wt.% K_2O , but most is in the range 0.2 to 1.0 wt.% K_2O (Fig. 8b), and it is variably enriched in barium, with up to 1.8 wt.% BaO , but most grains contain <0.5 wt.% BaO . The barium enrichment is noted to occur in olivine-free rocks. Plagioclase of composition $\text{An}_{6-14}\text{Ab}_{94-86}\text{Or}_{0-6}$ occurs within the symplectite areas of the granophyre sample (GGU-434207) and is intergrown with sanidine ($\text{Or}_{25-43}\text{Ab}_{75-57}$), quartz and apatite. In the area around the carbonate-filled cavity (Fig. 3a), the plagioclase is $\text{Ab}_{56-66}\text{Or}_{44-25}\text{An}_{0-10}$, which contrasts with the normal calcium-rich compositions in the rest of this sample (GGU-434023). In some cases, the plagioclase in areas proximal to carbonate alteration of mafic phases also gave unusual compositions ($\text{Ab}_{68-78}\text{Or}_{6-25}\text{An}_{7-20}$) with enrichment of Ba (to 2.5 wt.% BaO) and P (to 3.5 wt.% P_2O_5).

Orthopyroxene, occurring as subhedral grains in the matrix and around matrix olivine (Fig. 2e), has a variable composition, with $\text{Mg}/(\text{Mg} + \text{Fe})$ in the range 0.37 to 0.74, and with 1 to 2.5 wt.% CaO . This large variation may in part reflect the composition of the precursor olivine that it appears to have been replaced in some cases; it has a similarly large variation in $\text{Fe}:\text{Mg}$.

The Fe–Ti oxide phase consists of two distinct chemical phases, one of which has similar proportions of Fe and Ti (48 wt.% TiO_2 , 46–48 wt.% FeO) and one relatively enriched in Fe (67 wt.% FeO , 20 wt.% TiO_2 , 4 wt.% Al_2O_3 and 3 wt.% MgO).

Biotite occurs as a matrix phase in sample GGU-434009 and a phenocrystic phase in sample GGU-434207. The composition is uniform with $\text{Fe}/(\text{Fe} + \text{Mg})$ of 0.65 ± 0.02 , 4 to 5 wt.% TiO_2 , 25 wt.% FeO and 7 wt.% MgO (Table 5). The presence of annite in sample GGU-434207 relates to hydrothermal alteration (see below).

TABLE 3. REPRESENTATIVE COMPOSITIONS OF PYROXENE, NORTH GREENLAND DYKE ROCKS

Sample	GGU 434024	GGU 434024	GGU 434024	GGU 434-24	GGU 434017	GGU 434017	GGU 434017	GGU 434009	GGU 434009	GGU 434009	GGU 434009	GGU 434009
Point	1	17	24	25	1	2	3	2	6	7	17	18
Notes	core	core	core	rim	core	rim	core	matrix	core	rim	core	margin
SiO ₂ wt. %	48.38	44.80	48.70	45.61	48.76	50.79	48.46	51.44	48.3	51.38	50.26	49.81
TiO ₂	1.98	3.99	2.05	3.51	1.90	1.25	1.92	1.08	2.74	1.05	1.95	0.86
Al ₂ O ₃	3.26	6.32	3.14	5.61	3.80	1.99	3.50	1.59	4.32	1.66	3.7	2.17
FeO	9.30	9.58	9.72	10.13	9.99	10.96	10.14	15.71	11.46	14.51	9.82	17.24
MnO	0.14	0.10	0.19	0.19	0.22	0.16	0.21	0.53	0.00	0.42	0.00	0.45
MgO	14.46	12.21	14.25	12.27	14.31	14.76	13.82	14.75	13.29	14.87	14.97	10.42
CaO	20.73	21.46	20.33	21.29	19.32	19.20	19.39	15.41	19.46	16.54	19.76	19.22
Na ₂ O	0.48	0.67	0.40	0.59	0.50	0.48	1.14	0.3	0.48	0.32	0.28	0.47
Total	98.76	99.16	98.81	99.24	98.83	99.61	98.61	100.81	100.05	100.75	100.74	100.64
Structural formulae (6 atoms of oxygen)												
Si <i>apfu</i>	1.830	1.698	1.830	1.728	1.830	1.890	1.830	1.932	1.824	1.926	1.860	1.914
^{iv} Al	0.144	0.282	0.138	0.252	0.168	0.090	0.156	0.072	0.176	0.072	0.140	0.086
^{vi} Al	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.016	0.072	0.022	0.010
Ti	0.054	0.114	0.060	0.102	0.054	0.036	0.054	0.030	0.078	0.030	0.054	0.024
Fe	0.294	0.306	0.306	0.324	0.312	0.342	0.318	0.492	0.360	0.456	0.306	0.552
Mn	0.006	0.006	0.006	0.006	0.006	0.006	0.006	0.018	0.000	0.012	0.000	0.012
Mg	0.816	0.690	0.798	0.696	0.678	0.822	0.780	0.828	0.750	0.834	0.822	0.594
Ca	0.840	0.870	0.822	0.864	0.774	0.768	0.786	0.618	0.786	0.666	0.780	0.792
Na	0.036	0.048	0.030	0.042	0.036	0.036	0.084	0.024	0.036	0.024	0.018	0.036

The structural formulae are expressed in atoms per formula unit, *apfu*.

TABLE 4. REPRESENTATIVE COMPOSITIONS OF FELDSPAR, NORTH GREENLAND DYKE ROCKS

Sample	GGU 434024	GGU 434024	GGU 434009	GGU 434009	GGU 434009	GGU 434011	GGU 434011	GGU 434011	GGU 434011	GGU 434011	GGU 434009	GGU 434009	GGU 434009	GGU 434009
Point	14	15	3	28	1	5	22	3	4	16	17	18	19	21
Notes	core	rim	vesicle	vesicle	matrix	core	core	core	rim	core	rim	graphic	symp	symp
SiO ₂ wt. %	51.20	56.24	62.27	62.11	56.39	55.15	53.18	52.26	52.22	51.78	57.31	63.71	63.45	66.31
Al ₂ O ₃	29.72	26.79	21.27	23.99	27.32	28.20	28.88	29.38	29.56	29.96	26.34	22.92	23.09	22.07
CaO	13.44	9.08	1.88	5.54	9.96	11.16	12.50	12.90	12.95	13.46	8.87	3.96	4.67	2.73
Na ₂ O	3.85	5.77	6.16	7.60	5.75	5.22	4.38	4.17	4.12	3.96	5.98	7.53	7.62	7.69
K ₂ O	0.28	0.36	3.66	0.66	0.24	0.37	0.24	0.32	0.25	0.24	0.85	0.52	0.69	1.07
BaO	0.00	0.42	3.74	0.00	0.00	0.00	0.00	0.22	0.00	0.00	0.00	0.00	0.00	0.00
Total	98.49	98.66	98.98	99.90	99.66	100.10	99.18	99.24	99.09	99.40	99.34	98.64	99.52	99.87
An %	65.1	45.5	0.0	29.0	49.0	54.0	61.0	61.9	62.4	64.3	42.7	0.0	0.0	0.0
Ab %	33.3	52.1	11.1	68.1	50.4	44.9	38.1	36.5	36.0	34.1	52.4	74.3	71.3	78.1
Or %	1.6	2.5	63.9	4.2	1.6	2.4	1.6	1.6	1.6	1.6	4.8	3.7	4.4	6.7

Note: vesicle refers to feldspar lining vesicle walls; symp refers to feldspar in symplectitic texture. The proportion of feldspar end-members is expressed in mole %.

Composition of the secondary minerals

The secondary minerals occur in two associations, as alteration of Fe–Mg silicates (*e.g.*, olivine, pyroxene, biotite) and feldspar, or infilling voids, possibly primary

gas-escape structures (Fig. 3a, Table 6). The secondary minerals vary from trace amounts to abundant in samples, as reflected in LOI values for the whole rocks (up to 5 wt.%; see Table 6). In the first case, a variety of unidentified Fe–Mg–Al and Ti–Si–Al–Fe silicates oc-

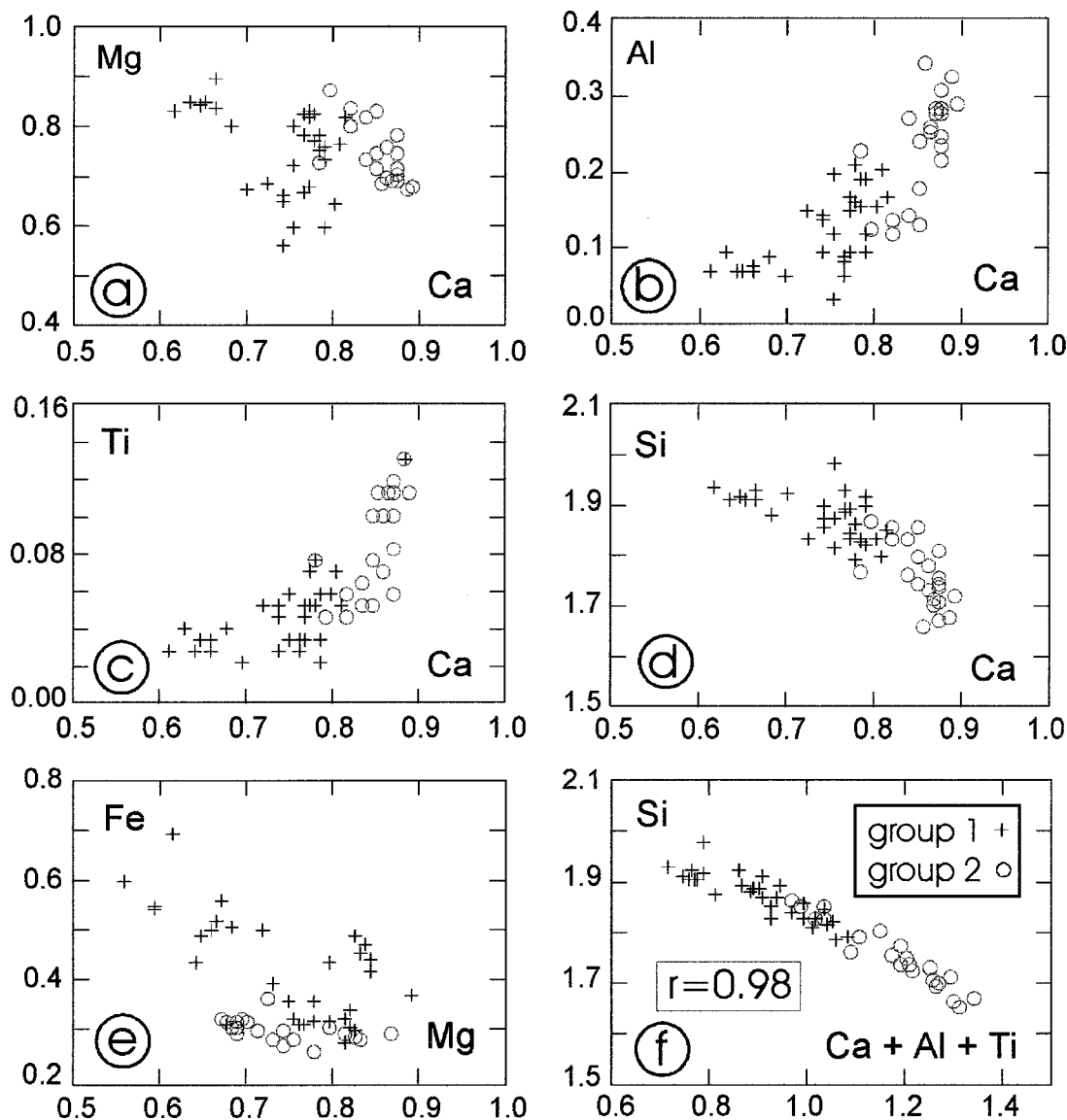


FIG. 7. Binary plots of the major cations in clinopyroxene (calculated on the basis of six atoms of oxygen) from the mafic dyke-rocks. Note the strong correlation of the data in the Si versus Ca + Al + Ti diagram.

cur after primary ferromagnesian minerals, whereas carbonate (manganiferous calcite, siderite, ankerite) and white mica (<2–3 wt.% FeO + MgO and trace TiO₂) occur after plagioclase. Potassium-rich feldspar (Or_{90–93}Ab_{7–10}) with up to 0.65 wt.% BaO occurs in the granophyre and may reflect re-equilibration of sanidine with a fluid phase. Secondary chlorite may show oscillatory zoning, with iron generally enriched in the core (33 wt.% FeO) versus the margin (28 wt.% FeO). Bi-

otite associated with chlorite is of annite composition, with Fe/(Fe + Mg) approaching 0.90.

Void areas occluded with secondary hydrothermal phases have a complex mineralogy (Figs. 2g, h, i, 3a), including a rim of quartz, feldspar (Or_{25–43}Ab_{66–56}An_{0–11}) and hydrated Fe–Mg–Al silicates, which surround coarse calcite with trace to minor amounts of Fe and Mn.

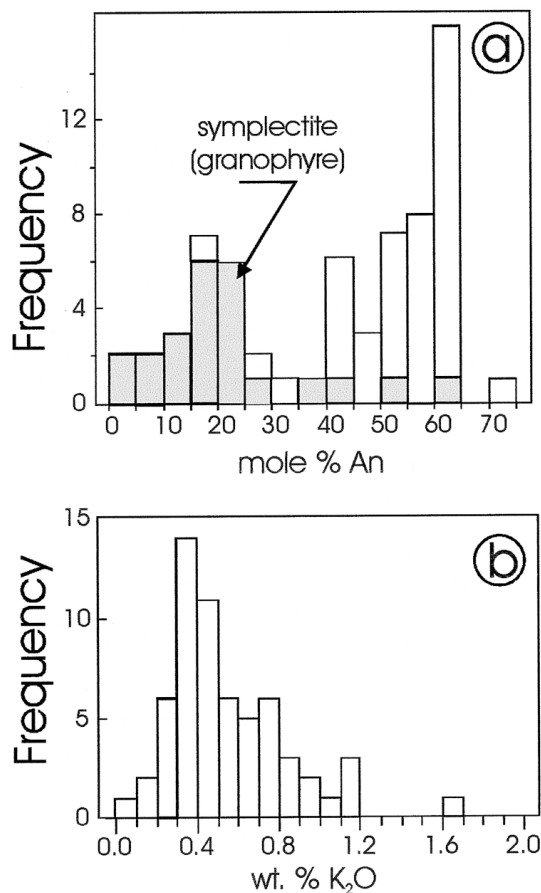


FIG. 8. Histogram plots summarizing chemical features of plagioclase in mafic dyke-rocks. (a) Plot of mole % An for core and rim of plagioclase grains for all groupings of dyke rocks, rim compositions for plagioclase in sample GGU-434024 (dark pattern) and the sodic plagioclase (dark pattern and compositions $<An_{20}$) in the granophytic areas of sample GGU-434207 (e.g., Fig. 2f). (b) Plot of wt.% K_2O in plagioclase grains for all groupings of dyke rocks.

Whole-rock geochemistry: major elements

The eight samples of dyke rock analyzed (Table 7) can be subdivided into three separate groups that chemically correspond to the groupings (1, 2, 3) discussed previously.

The rocks are of basaltic composition in terms of their silica content (ca. 47 wt.% SiO_2 on LOI-free basis), with the low silica contents of some samples (e.g., GGU-434022, 23, 24) corresponding to picritic basalts (Le Maitre *et al.* 1989). The combined high LOI and CO_2 (to 6.6 wt.% in GGU-434024) contents in some of the rocks reflect the presence of calcite occluding vugs,

TABLE 5. REPRESENTATIVE COMPOSITIONS OF BIOTITE, NORTH GREENLAND DYKE ROCKS

Sample	GGU 434009	GGU 434009	GGU 434009	GGU 434009	GGU 434009	GGU 434009	GGU 434207	GGU 434207	GGU 434207
Point	12	25	1	3	5	6	10	11	2
SiO_2 wt. %	35.48	35.12	36.09	35.93	35.60	35.80	32.76	33.43	35.33
TiO_2	5.03	4.89	3.22	4.41	5.08	5.05	4.26	4.68	4.40
Al_2O_3	12.53	12.42	11.89	12.22	12.32	12.41	16.40	16.08	14.53
FeO	26.27	26.98	26.60	25.69	25.62	25.65	30.44	31.56	26.26
MgO	7.18	6.77	8.14	8.05	7.74	7.61	2.27	1.51	6.36
CaO	0.35	0.80	0.00	0.00	0.00	0.22	0.00	0.00	0.00
Na_2O	0.00	0.00	0.37	0.31	0.65	0.36	0.00	0.00	0.00
K_2O	8.72	8.77	8.55	8.76	8.54	8.90	8.38	8.50	8.32
Cl	0.16	0.21	0.31	0.29	0.42	0.40	0.00	0.00	0.00
Total	95.72	95.96	95.17	95.66	95.97	96.40	94.51	95.76	95.20
Structural formulae (11 atoms of oxygen)									
Si <i>apfu</i>	2.794	2.761	2.871	2.838	2.805	2.805	2.673	2.695	2.772
^{IV}Al	1.206	1.239	1.129	1.162	1.195	1.195	1.327	1.305	1.228
^{VI}Al	0.000	0.000	0.000	0.000	0.000	0.000	0.246	0.224	0.114
Ti	0.297	0.286	0.198	0.264	0.297	0.297	0.264	0.286	0.264
Fe	1.727	1.771	1.760	1.694	1.683	1.683	2.068	2.123	1.727
Mn	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.022	0.022
Mg	0.847	0.792	0.968	0.946	0.913	0.891	0.275	0.187	0.748
Ca	0.033	0.066	0.000	0.000	0.000	0.022	0.000	0.000	0.000
Na	0.000	0.000	0.055	0.044	0.099	0.055	0.000	0.000	0.000
K	0.880	0.880	0.869	0.880	0.858	0.891	0.869	0.869	0.836
Cl	0.022	0.033	0.044	0.044	0.055	0.055	0.000	0.000	0.000
Fe/(Fe + Mg)	0.67	0.69	0.65	0.64	0.65	0.65	0.88	0.92	0.69

The values of Fe/(Fe + Mg) show that the dark mica is annite.

the related alteration of mafic minerals, and formation of secondary hydrated Fe-Mg silicates near these areas (see above, Figs. 2g, h, i). All the rocks show notable enrichment in FeO_T , P_2O_5 , TiO_2 and K_2O ; the least-altered rocks (GGU-434009, 11, 17, 19) fall on the boundary between basanite and basalts in the silica versus $Na_2O + K_2O$ plot of Le Maitre *et al.* (1989). Group-1 rocks are relatively enriched in FeO_T , MgO and Al_2O_3 , and show less enrichment in P_2O_5 , LOI and CO_2 , whereas group-2 rocks are relatively enriched in P_2O_5 , LOI, CO_2 , K_2O and Na_2O , consistent with the increased abundance of plagioclase, apatite, and variably developed carbonate alteration in these rocks. The final sample (GGU-434207), which is characterized by the presence of the granophyre texture and abundant secondary minerals, is relatively enriched in SiO_2 (51.7 wt.%) and Al_2O_3 , and strongly depleted in TiO_2 and FeO_T and CaO, with minor depletion in MgO and P_2O_5 .

The normative mineralogy of the dyke rocks, calculated after adjusting the $FeO:Fe_2O_3$ ratio (Irvine & Baragar 1971), also corresponds to the three groupings mentioned above. Group-1 rocks are Hy-normative, whereas group-2 rocks lack Hy and instead have trace Ne, and both group-1 and 2 rocks are Ol-normative. The single group-3 rock is quartz-normative, which may reflect in part the high degree of alteration. Thus, the dyke rocks can broadly be grouped on the basis of their

TABLE 6. REPRESENTATIVE COMPOSITION OF ALTERATION MINERALS, NORTH GREENLAND DYKE ROCKS

Sample	GGU 434009	GGU 434009	GGU 434009	GGU 434009	GGU 434011	GGU 434011	GGU 434011	GGU 434019	GGU 434027
SiO ₂ wt. %	44.08	48.86	49.32	41.43	47.22	43.47	49.33	45.12	44.65
TiO ₂	0.00	0.00	0.00	0.00	0.27	0.12	0.10	0.00	0.00
Al ₂ O ₃	3.05	2.05	0.00	0.66	3.66	0.18	0.14	4.69	4.54
FeO	32.78	20.08	21.18	25.67	26.02	20.02	12.71	23.48	23.68
MnO	0.97	0.00	0.26	0.00	0.00	0.16	0.00	0.00	0.31
MgO	2.90	17.99	21.43	16.50	4.50	22.67	25.68	11.01	10.23
CaO	1.61	0.51	0.29	2.80	1.03	0.30	0.16	1.65	1.80
Na ₂ O	0.00	0.00	0.29	0.00	0.48	0.63	0.32	1.03	0.00
K ₂ O	0.00	0.00	0.00	0.00	0.54	0.05	0.07	0.42	0.66
Total	85.39	89.49	92.77	87.06	83.71	87.59	88.51	87.40	85.87
Sample	GGU 434023	GGU 434023	GGU 434023	GGU 434023	GGU 434024	GGU 434024	GGU 434024	GGU 434024	GGU 434024
SiO ₂ wt. %	40.67	33.28	38.27	46.65	0.14	0.70	0.24	25.67	28.58
TiO ₂	0.00	0.00	0.00	0.00	0.00	0.00	0.00	56.86	47.60
Al ₂ O ₃	4.96	4.34	4.66	4.90	38.09	39.31	28.80	3.40	3.73
FeO	15.87	13.83	15.31	22.51	0.46	0.45	0.28	3.98	6.65
MnO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
MgO	13.92	9.88	12.82	10.79	8.99	8.35	18.41	1.37	2.77
CaO	1.18	1.25	1.15	1.28	7.21	6.17	6.09	1.43	1.44
Na ₂ O	0.62	0.54	0.59	1.12	0.32	0.33	0.37	0.84	0.84
K ₂ O	0.52	0.51	0.48	1.19	0.00	0.00	0.05	0.41	0.00
Total	77.74	63.63	73.28	88.44	55.22	55.31	54.23	93.96	91.61

normative mineralogy as olivine tholeiite (group 1), alkali basalt (group 2) and basaltic andesite (group 3).

Whole-rock geochemistry: trace elements

The trace element abundances of the dyke rocks (Fig. 9, Table 7) correlate with the groupings noted (e.g., Zr versus Ba, Cr versus V; Figs. 9b and 9f, respectively). In general, group-1 rocks are relatively enriched in V, Cr, Cu, Ni, Co, Zr, Pb and depleted in Ba and Sr compared to group-2 rocks, but note that sample GGU-434019 is depleted in Cr, Ni and Cu compared to other group-1 rocks. Similar levels of enrichment occur for Sc, Rb, Ga, Nb and Th; both groups have levels of U, Mo and Sn near the detection limits. The last sample (GGU-434207) shows irregular enrichment or depletion; even though it has Ni values intermediate between groups 1 and 2, its Ba and Sr contents are comparable to group 1.

Chondrite-normalized rare-earth-element (REE) plots (Fig. 10) show that all samples are similar, with enrichment of Σ REE and strongly fractionated patterns, with enrichment in the light REE (LREE) [i.e., (La/Lu)_N values about 10]. Such patterns are typical of alkaline basalts in general (e.g., Cullers & Graf 1984, Sun & McDonough 1989). Group-1 rocks are slightly depleted in the LREE and enriched in the heavy REE (HREE), with only a slight positive Eu anomaly compared to group-2 rocks, which have a markedly positive Eu anomaly. Sample GGU-434207 shows the strongest LREE enrichment and HREE depletion, and has no Eu

TABLE 7. GEOCHEMISTRY OF DYKE ROCKS, NORTH GREENLAND

Sample	GGU 434009	GGU 434011	GGU 43017	GGU 434019	GGU 434022	GGU 434023	GGU 434024	GGU 434207
Location	1	1	1	2	2	2	2	1
SiO ₂ wt. %	47.18	47.31	47.25	46.80	43.83	45.67	41.98	51.78
TiO ₂	4.20	4.15	4.16	4.17	4.02	4.12	3.77	2.32
Al ₂ O ₃	13.94	13.95	13.96	13.17	13.53	14.27	13.09	17.04
Fe ₂ O ₃	2.83	0.09	3.61	2.06	3.74	4.09	0.00	0.00
FeO	11.60	13.93	11.11	12.22	8.85	8.92	13.29	12.28
Mn	0.23	0.23	0.23	0.22	0.21	0.23	0.22	0.19
MgO	5.63	5.67	5.79	5.37	5.06	4.74	5.31	4.52
CaO	9.80	9.70	9.71	9.13	9.20	9.48	9.72	3.58
Na ₂ O	2.82	2.69	2.93	2.85	3.23	3.43	2.93	3.39
K ₂ O	0.90	0.97	0.83	1.14	1.03	1.22	1.00	0.98
P ₂ O ₅	0.75	0.74	0.75	0.87	1.37	1.23	1.07	0.63
Cl	0.03	0.02	0.03	0.01	0.01	0.00	0.01	0.01
CO ₂	0.16	0.36	0.06	0.10	3.00	1.54	6.64	0.55
H ₂ O _T	0.41	0.13	0.14	1.32	1.67	3.21	0.47	3.19
¹⁴³ Nd/ ¹⁴⁴ Nd	0.51285	0.51282						
⁸⁷ Sr/ ⁸⁶ Sr	0.70496	0.70471						
δ ¹⁸ O ‰	7.6	10.5	7.9	7.2	7.0	6.0	6.7	9.9
Mg#	0.44	0.45	0.44	0.43	0.44	0.42	0.45	0.43
Trace elements (XRF except for *, which are ICP-MS)								
Rb ppm	35	33	33	41	26	30	26	33
Sr	473	473	499	535	565	537	530	422
Ba	464	549	706	573	2276	2071	2375	531
Li *	14.3	6.1	9.2	9.7	12.8	7.6	11.0	178.9
Cs *	1.5	2.2	0.9	0.5	0.2	0.1	1.1	1.4
Ga	13	13	11	12	13	13	12	17
Nb	29	29	26	29	28	27	25	20
Sn	ND	2	1	3	3	ND	ND	2
Ni	44	45	37	ND	ND	ND	6	27
Cu	100	102	104	24	28	31	32	58
Zn	118	106	99	103	99	78	95	167
Pb	6	6	8	5	2	2	2	24
Th *	2.8	2.7	2.7	3.3	2.1	2.4	1.8	6.8
U *	0.9	0.9	0.8	1.0	0.6	0.7	0.5	0.8
Zr	208	211	236	226	164	169	175	138
Y *	32.7	30.6	31.5	33.3	30.9	29.1	27.8	23.5
Hf *	5.6	5.2	5.5	5.3	3.9	4.1	3.4	1.6
Ta *	2.0	2.0	2.0	2.1	2.0	2.0	1.7	1.3
Sc	4	3	2	4	7	10	ND	12
V	341	330	339	388	272	283	300	163
Cr	83	74	73	39	15	18	44	9
Co	47	44	45	43	31	32	33	41
Rare-earth elements								
La ppm	28.01	27.78	27.91	30.33	32.13	30.82	26.11	36.69
Ce	65.36	64.20	64.05	67.70	72.90	68.77	59.95	83.37
Pr	8.75	8.52	8.47	9.08	10.09	9.25	8.31	10.22
Nd	37.62	36.98	36.63	39.91	45.66	42.51	37.75	41.41
Sm	8.78	8.46	8.55	8.92	10.09	9.35	8.19	8.43
Eu	3.29	3.02	3.21	3.43	6.35	5.91	5.21	2.52
Gd	8.45	8.00	8.32	9.04	10.40	8.97	8.13	7.47
Tb	1.26	1.19	1.19	1.23	1.32	1.19	1.11	1.01
Dy	7.07	6.87	7.17	7.38	7.26	6.83	5.93	5.57
Ho	1.37	1.30	1.34	1.36	1.30	1.18	1.12	0.99
Er	3.49	3.29	3.44	3.62	3.24	3.10	2.87	2.50
Tm	0.49	0.46	0.51	0.50	0.41	0.41	0.37	0.32
Yb	2.96	2.72	2.88	2.96	2.41	2.24	2.17	1.71
Lu	0.42	0.41	0.41	0.45	0.35	0.35	0.30	0.23

Mg# = Mg/[Mg + Fe(total)]. Locations: 1 Frigg Fjord, 2 Harebugt.

anomaly. Thus, there is a general rotation of the REE data from the most primitive to evolved rocks such that LREE enrichment and HREE depletion occur. The rep-

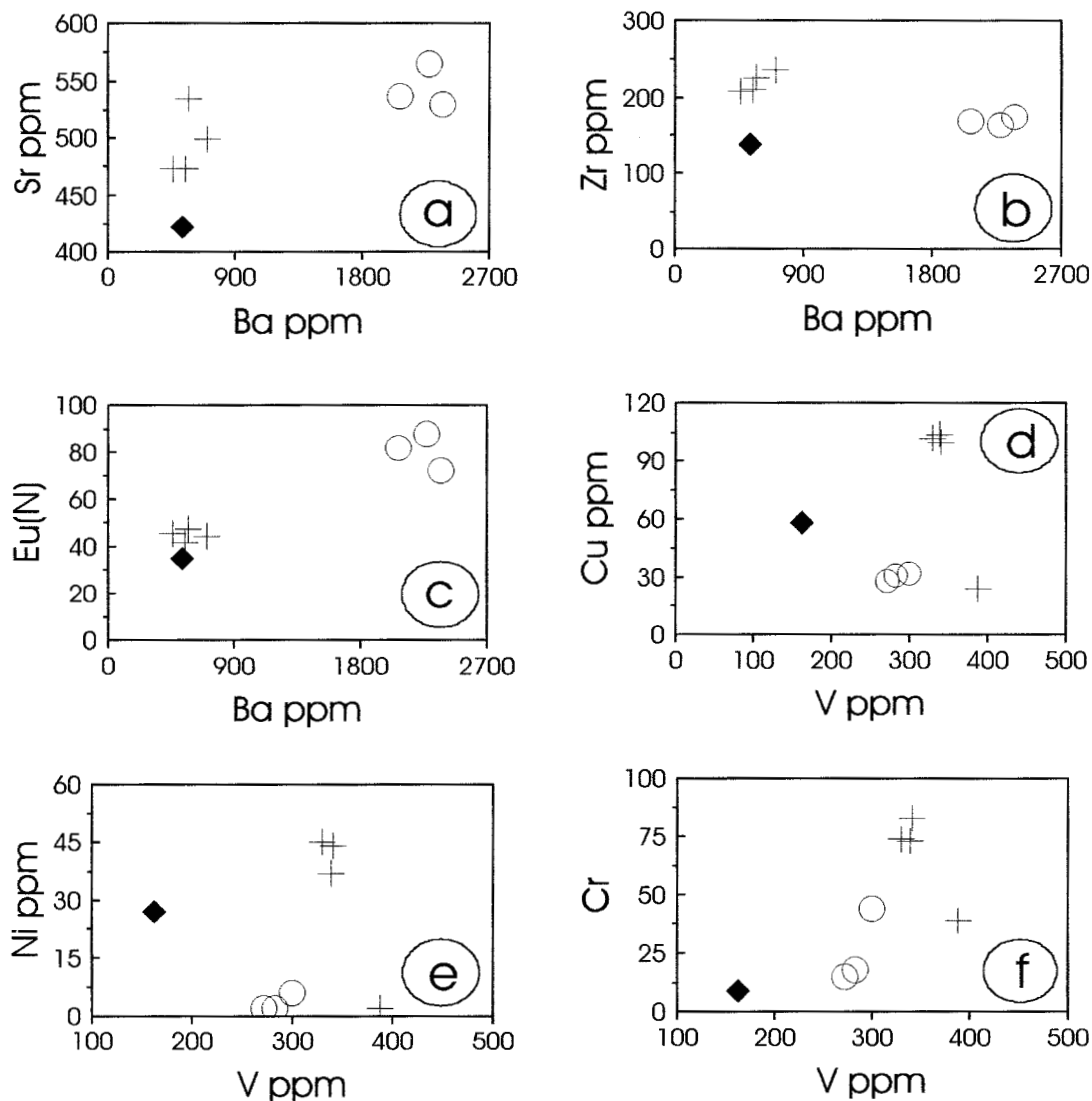


FIG. 9. Trace-element geochemistry of mafic dyke rocks from north Greenland showing the three groupings, as discussed in the text (+ group 1, ○ group 2, ◆ group 3). The relative scatter of data for sample GGU-434207 (diamond symbol) reflects the presence of alteration in this sample (*e.g.*, Ni, Cu enrichment, but Ba depletion and low Eu_N value).

licate analysis of sample GGU-434207 indicates excellent reproducibility of the data.

Stable (O, C) and radiogenic (Sr, Nd) isotopes

All whole-rock samples were analyzed in duplicate for $\delta^{18}\text{O}$ (Table 7), and results range from +6.0 to +10.5‰. There is no apparent correlation of the data with either petrographic or geochemical features of the samples that might explain some of the elevated $\delta^{18}\text{O}$

values. For example, although GGU-434207 is chemically the most evolved sample and is altered, sample GGU-434011 is petrographically the freshest sample analyzed and yet has the highest $\delta^{18}\text{O}$ value. The range is well outside that attributed to crystal fractionation, even excluding the two most enriched samples (Taylor & Sheppard 1986). If the two elevated samples are ignored, the remaining samples with $\delta^{18}\text{O}$ of $7.1 \pm 0.6\text{‰}$ compare to the global range of $\delta^{18}\text{O}$ values for alkaline basalts and basic lavas (Kyser 1986).

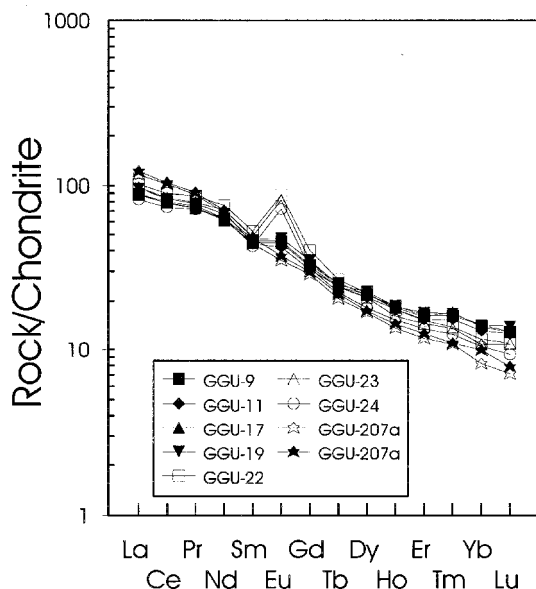


FIG. 10. Chondrite-normalized rare-earth-element plots (normalizing factors of Masuda *et al.* 1973) for mafic dyke-rocks from North Greenland. Note that the samples have been subdivided into the three groupings and that group-2 dyke rocks (GGU-434022, 023, 024) are characterized by relatively depleted *HREE* and a more pronounced positive Eu anomaly compared to the other samples of dyke rock. Results of duplicate analyses of sample GGU-434207a are shown.

The $^{87}\text{Sr}/^{86}\text{Sr}$ values measured in two whole-rock samples (Table 7) are similar at $0.70496 (\pm 0.00004)$ and $0.70471 (\pm 0.00003)$; the corrected $\text{Sr}_i(85 \text{ Ma})$ values of 0.70477 and 0.70458 , respectively, correspond in general to those expected of fresh basaltic rocks. The same two dyke-rocks were also analyzed for $^{143}\text{Nd}/^{144}\text{Nd}$, and the results are essentially identical at $ca. 0.51283 \pm 0.00002$, with corrected initial ratios of 0.51275 . Model Nd–Sm ages for the samples calculated for depleted mantle (T_{DM}) evolution are $ca. 720 \text{ Ma}$. In terms of a plot of $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ (Hart 1988) (Fig. 11a), which is used to classify oceanic basalts, the dyke samples plot within the OIB field, but not close to any of the end members (*i.e.*, well away from EM I, EM II and HIMU).

Three samples of transparent calcite spar filling primary vesicles or ocelli in the core of a dyke rock (GGU-434023) indicate uniformity for both $\delta^{18}\text{O}$ (+15.7, +15.5, +15.6‰) and $\delta^{13}\text{C}$ (–5.5, –5.8, –5.8‰). Calculated $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ values, assuming equilibrium between fluid and calcite, are +13 to +14‰ for the range 400–500°C (equation of O'Neil *et al.* 1969), which are considered reasonable values for a cooling mafic intrusion

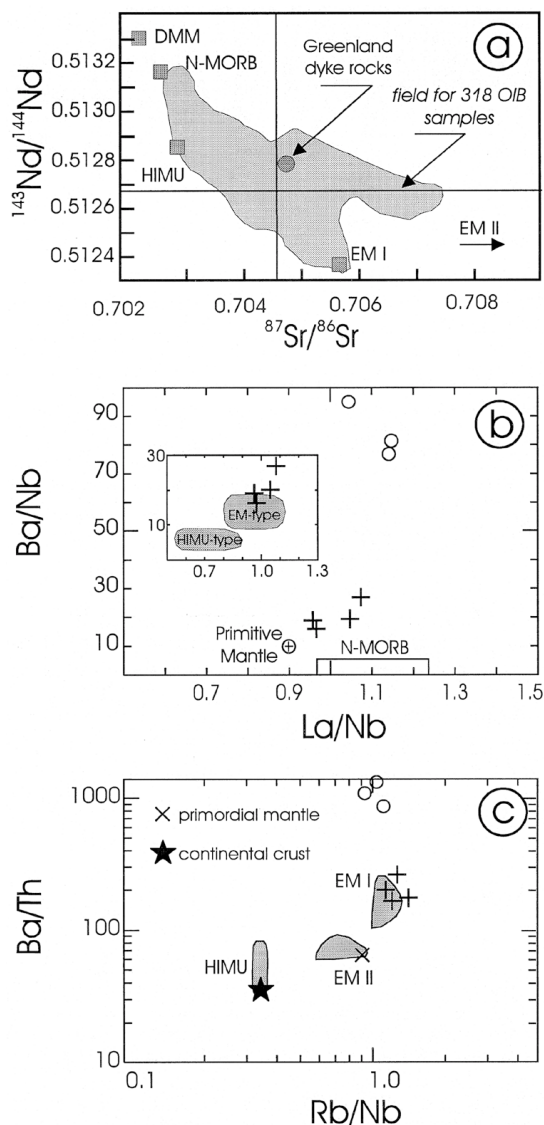


FIG. 11. (a) Plot of $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ for Greenland dyke-rocks ($n = 2$, Table 2) compared to field for Ocean Island basalts (OIB) and mantle reservoirs of Hart (1988) (DMM = depleted MORB mantle); straight lines represent bulk-earth isotopic values (Zindler & Hart 1986). (b) Binary plot of Ba/Nb versus La/Nb (after Sun & McDonough 1989) for Greenland dyke-rock samples compared to fields for HIMU- and EM-type basalts (shown in the inset). Note the strong positive correlation among the samples. (c) Binary plot (log units) of Ba/Th versus Rb/Nb for Greenland dyke-rock samples compared to fields for HIMU, EM I and EM II reservoirs (after Weaver 1991).

(note that $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ values increase moderately for higher temperatures: at 600°C , $\delta^{18}\text{O}_{\text{H}_2\text{O}} = +16.8\text{‰}$). These $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ values are well outside the range for magmatic water (*i.e.*, $+6$ to $+10\text{‰}$; Taylor 1979, Sheppard 1986) and suggest an incursion of an externally derived fluid or equilibration of a magmatic fluid with the metasedimentary wallrocks. The $\delta^{13}\text{C}$ values overlap the fields for magmatic and metamorphic carbon (Ohmoto & Rye 1979); hence, the data are equivocal in terms of a possible source of carbon.

DISCUSSION

Significance of a Cretaceous age for the dyke rocks

The best estimate for emplacement of the dyke rocks is provided by the 81.9 ± 1.0 Ma plateau age for sample GGU-434019 from Harebugt, which agrees within error with the isotope correlation age of 85.6 ± 4.7 (magnetic and nonmagnetic fractions) for sample GGU-434009 and the resetting of the hydrothermally altered metasediment sample GGU-434016 at 80 ± 6 Ma, also from Frigg Fjord. We adopt 82 Ma as the best estimate for the timing of dyke emplacement; the dates indicate that dyke emplacement at the two localities along Frederick E. Hyde Fjord (Fig. 1) took place at the same time. The ages obtained in this study are somewhat older than the previously preferred whole-rock K-Ar ages of 72 to 66 Ma for the dyke rocks (Soper *et al.* 1982, Dawes *et al.* 1983), and are also some 20 Ma older than the well-constrained timing of bimodal peralkaline volcanism represented by the Kap Washington Group (Larsen *et al.* 1978, Larsen 1982, Batten 1982). In addition, the *ca.* 82 Ma age for the dykes is also some 25 Ma older than the inferred timing of dyke emplacement for the southern suite of dyke rocks based on the paleomagnetic measurements of Abrahamsen *et al.* (1997).

The relevance of these new ages relates to the paleotectonic configuration of northern Greenland at this time and the nature of magmatism occurring within the area. The 82 Ma age for the dyke rocks is wholly consistent with regional extension and with a cause-and-effect relationship between dyke emplacement and formation of the Makarov Basin at this time, which is best constrained to have formed at anomaly 34 time (*i.e.*, Santonian, 83 to 86 Ma; Taylor *et al.* 1981). This conclusion is consistent with the observation of Soper *et al.* (1982), that the orientation of the dyke rocks agrees better with the orientation of the Makarov spreading axis than the Nansen Ridge, which was not a feature until anomaly 24 time (52 Ma). Thus, the dyke rocks would have formed within a continental setting and, in fact, represent the initial stages of what would subsequently become a continental rift. The tentative suggestion, discussed below, that the dyke rocks are a surface manifestation of a mantle plume raises the possibility that formation of the Makarov Basin was in fact initiated by this plume.

Although Soper *et al.* (1982) and Soper & Higgins (1991) discussed the possible relationship between the dyke swarm and extension related to formation of the Makarov Basin, they expressed concerns over this alkaline magmatism and the generally contemporaneous tholeiitic magmatism exposed on adjacent Ellesmere Island and the Barents margin of the Eurasia basin in Svalbard and Franz Joseph Land [Tyrell & Sandford 1933, Gusev 1979 (in Soper *et al.* 1982)]. However, petrological studies have demonstrated that in many cases, alkaline-type plume-related magmatism gives way laterally to tholeiitic suites. In addition, the spatial and temporal association in rift basins of alkaline, tholeiitic and peralkaline suites has also been demonstrated (Ramberg & Larsen 1978, Storey *et al.* 1999). Thus, formation of the peralkaline Kap Washington volcanic suite may represent subsequent generation of magma related to the same plume.

On a more regional scale, the age dates are consistent with the recent suggestion of Maher (2001) that the high Arctic large igneous province (HALIP) spanned much of the Lower and early part of the Upper Cretaceous. Magmatism of a similar age as that presented here is found in northern Ellesmere Island and Franz Josef Land; references are provided by Maher (2001).

Implications of partially reset ages in whole-rock metasedimentary rocks

The age of the low-temperature gas fractions for the samples of metasedimentary rock suggests that these samples were affected by a low-temperature overprinting thermal event. The age of *ca.* 80 Ma for the low-temperature gas fraction released in sample GGU-434016 relates this resetting to intrusion of the dyke rocks. This resetting of the age spectra for the metasedimentary rocks was unexpected, given the generally narrow width (*i.e.*, <10 – 15 m) of the dykes, the rapid loss of heat in such narrow dykes to the surrounding country-rocks *via* conduction, and the rarity of the dykes within the study area. Of particular relevance here is the fact that sample GGU-434016 contains abundant white mica and is extremely altered, possibly related to fluid circulation initiated by emplacement of the dyke rocks. The resetting recorded by the metasedimentary rocks may indicate that a larger volume of mafic dyke material is located at depth, *i.e.*, the exposures represent the tip of a much larger dyke-swarm.

Nature and source of the magma that led to the dyke rocks

Petrographic features of the dykes indicate an absence of coarse phenocrystic phases and minor microphenocrysts of clinopyroxene in an ophitic texture with plagioclase and rare glomeroclasts of the same minerals. Thus, the dyke rocks are considered to approximate liquid compositions. The olivine composi-

tions (Fo_{69–35}) combined with the Ni contents (<40 ppm) and low values of mg# (0.42 to 0.45) suggest extended fractionation of the melts compared to primitive magmas [e.g., Ni in the range 235–400 ppm (Sato 1977), Mg/(Mg + Fe²⁺) in the range 0.63–0.73 (Green 1971)]. Separation of the melt from early-crystallizing phases may have occurred as a result of flowage differentiation, given the absence of significant phenocrystic phases. Thus, the dyke rocks probably represent products of fractionation of a less evolved, relatively primitive parental magma, the source of which may reside at a considerable distance from the dykes, given that magma may travel great lateral distances in dyke swarms (Greenough & Hodych 1990, Ernst *et al.* 1995). The fact that plagioclase was not an early-crystallizing phase, as supported by the lack of a negative Eu anomaly and absence of plagioclase phenocrysts, is consistent with crystallization of a hydrous basaltic melt in which plagioclase nucleation is suppressed by an elevated H₂O content (e.g., Blatter & Carmichael 1998). The hydrous nature of the melt is also suggested by: (1) the inferred ocelli-like features (Fig. 3a), (2) presence of late annite in the matrix of some dykes and as a phenocrystic phase in the most evolved sample, and (3) late hydrothermal alteration, notable within the alkaline dykes (*i.e.*, group-2 dyke rocks).

As noted above, the petrographic features and normative mineralogy of the dyke rocks suggest that they can be subdivided into olivine tholeiites and alkali basalts, an association that is, of course, not unusual. Both rock types are well represented, for example, in Hawai'i, both temporally and spatially (Carmichael *et al.* 1974). We do, however, emphasize that the greater amount of alteration in group-2 dyke rocks may influence the classification of the latter on the basis of normative mineralogy, but at the same time note that the abundances and chondrite-normalized patterns of the REE, which are generally considered immobile and less affected by alteration, are very similar for both groups and thus suggest a similar petrogenesis. The chemical nature of the dyke rocks is further evaluated with the use of standard classification diagrams below. Incorporated in this discussion are the dyke rocks sampled by Soper *et al.* (1982). All data have been recast to 100% anhydrous prior to plotting to offset any effects of alteration.

In terms of major elements, the composition of the dykes partly overlaps the division between subalkaline and alkaline rocks in (Na₂O + K₂O) *versus* SiO₂ and Ol'-Ne'Q' plots of Irvine & Baragar (1971) (Figs. 12a, b) and are neither potassic or sodic, as illustrated in the An-Ab'-Or plot (Fig. 12c, also after Irvine & Baragar 1971). We also note that the data of Soper *et al.* (1982) show similar chemical characteristics in all these diagrams (Fig. 12). Thus, the two datasets compare well and indicate that there was similar chemical variation of the dyke-producing magmas over a large area. Dostal & Durning (1998) also noted chemical variation in their

compilation of data for a geographically comparable set of dyke samples, the Eastern North America (ENA) dyke swarm.

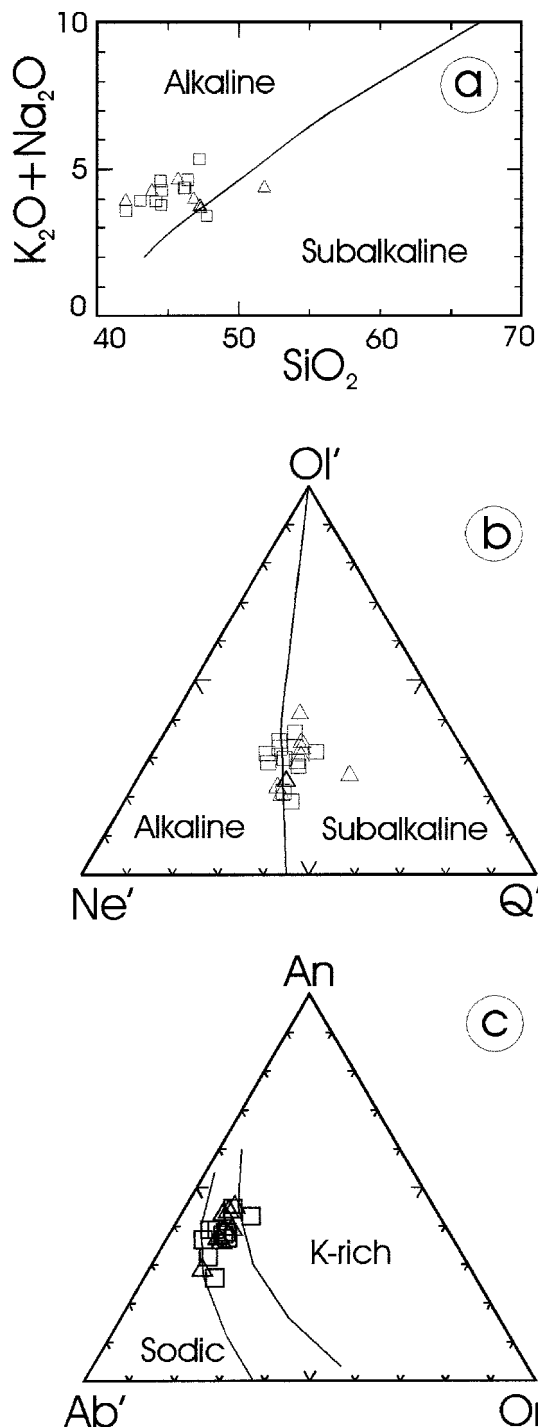
The ratios of incompatible elements have been used to infer the nature of the source regions of basalts. For example, the Th/Yb *versus* Ta/Yb plot of Pearce (1983) shows that the batches of dyke-producing magma fall in the mantle-array field and that the source area has not been influenced by either subduction-related enrichment or crustal contamination. The lack of subduction-related enrichment is further indicated by the absence of a negative Nb-Ta anomaly (Fig. 13a), the general presence of which infers a prior component of suprasubduction [see Kerrich & Wyman (1996) for recent review]. The lack of significant crustal contamination is supported by low Th/La (0.1), Th/Ta (1 to 1.5), Nb/Ta (15), and La/Ta (14) ratios, all sensitive indicators of crustal interaction (Green 1971, Green 1995, Sun & McDonough 1989, Taylor & McLennan 1985, Thompson & Morrison 1988). Although the low initial ⁸⁷Sr/⁸⁶Sr values suggest little involvement of the crust, the whole-rock $\delta^{18}\text{O}$ values and $\delta^{18}\text{O}_{\text{H}_2\text{O}}$ for ocelli-like carbonate in one dyke do indicate some crustal influence. We interpret these latter data to indicate probable incursion of an external fluid buffered by the metasedimentary wallrocks, as the trace-element chemistry of samples is uniform regardless of their $\delta^{18}\text{O}$ values. These conclusions, combined with the LREE and LILE enrichment of the dyke rocks, suggest that the geochemistry of the dyke-rock samples relate to the source region.

The type of the source region can be evaluated using the REE data and the element-ratio diagrams (e.g., La/Sm *versus* La, Sm/Yb *versus* Sm) of Aldanmaz *et al.* (2000), which consider PM (Primitive Mantle) and DMM (Depleted MORB Mantle) as global reservoirs. The LREE concentrations of the dyke rocks cannot be produced by even small amounts of partial melting of these sources, but instead about 5% partial melting of a spinel-garnet (50:50) lherzolite source can produce melts of the appropriate LREE enrichment. The elevated Ti/Yb values (Fig. 13) also indicate that garnet was a residual phase in the source region (Pearce 1996), for which a minimal estimate of the depth of melt generation would be ca. 60–80 km (Watson & McKenzie 1991). The elevated levels of K, Ti and Sr in the dyke-producing magma would indicate that the source contained phlogopite and pargasitic amphibole, presumably present as vein material within the lherzolite, and that no major Ti-bearing phase remained as a residual phase (Greenough 1988).

Tectonic setting of the dyke-producing magma

The data fall in the continental basalt field in the TiO₂-K₂O-P₂O₅ diagram of Pearce *et al.* (1975) (not shown) and the TiO₂-FeO_T/MgO plot of Fodor & Vetter (1984; Fig. 14a). In trace-element plots, the chemical data generally fall in the within-plate basalt fields

(Figs. 14b, c, e), although the data overlap the MORB–WPB field in the Th–Hf–Ta diagram of Wood *et al.* (1979) (not shown). The alkaline character is also indi-



cated in the Ti–V plot (Fig. 14d), which is mainly a reflection of the elevated TiO_2 contents of the samples. This feature is similarly illustrated by the fact that samples correspond to the OIB field in Eby's (1992) plots of Yb/Ta and Ce/Nb versus Y/Nb (not shown). Finally, the composition of the clinopyroxene indicates a transitional, anorogenic affinity for the dykes (Fig. 15). In this latter diagram, the separation of group-1 and -2 dyke rocks is clearly apparent.

An OIB affiliation for the dyke-producing magma

In Figure 13a, the compositions have been normalized to the primitive mantle of Sun & McDonough (1989). We emphasize the strongly fractionated nature of the dyke magma; note that sample GGU-434207 is omitted in Figure 13a because of its altered nature. Highlighted in Figure 13a is the generally uniform composition of the samples, but with small and consistent overall differences between group-1 and -2 samples (*e.g.*, Ba, Th, Pb, P, Zr, Hf, Eu, HREE). The fact that the same samples show relative enrichment in both Cs and Pb may relate to crustal contamination, for which we already noted support based on isotopic data. However, as further evidence precluding a significant role of crustal contamination, we compare the data of a representative sample (GGU-434009) to continental crust in Figure 13b and note the relative difference between the patterns.

Because of the similarities between the trace-element geochemistry of the dyke rocks and that of OIB rocks, a representative sample (GGU-434009) of fresh dyke rock with almost no phryic component is compared to the average OIB of Sun & McDonough (1989) in Figure 13b. They are very similar, except for Cs, Pb and Li. Thus, the dyke rocks have a clear affinity to the melts and processes to which OIB are related.

The compositional ranges of oceanic basalts, as established from radiogenic isotope ratios and trace elements, are considered to result from mixing of at least four end-members, namely depleted MORB mantle (DMM), HIMU mantle, and two enriched mantle components referred to as EM I and EM II (Zindler & Hart 1986, Hart 1988, Weaver 1991). Based on the type of plots used by Hart (1988) (Fig. 11a), the dyke-rock data fall between the HIMU and EM fields, with the latter

FIG. 12. Plots used for the chemical classification of mafic dyke-rocks from northern Greenland. Triangles are samples from this study, and boxes represent previously analyzed samples of dykes rocks for the same swarm located further north of the present study-area (data in Soper *et al.* 1982). All plots are inspired from Irvine & Baragar (1971): (a) $K_2O + Na_2O$ versus SiO_2 , (b) normative Ol' – Ne' – Q' plot to discriminate subalkaline from alkaline rocks, and (c) normative An – Ab' – Or plot to discriminate between sodic and potassic subalkaline suites.

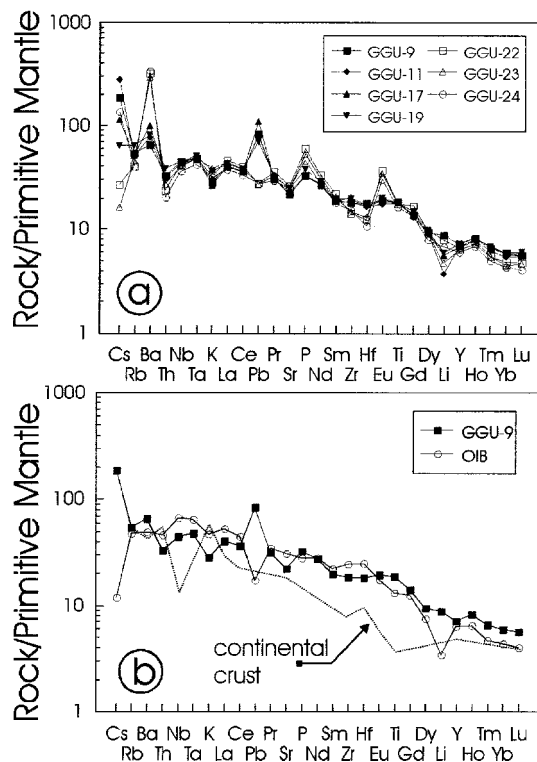


FIG. 13. (a) Extended trace-element normalized diagram for whole-rock samples of mafic dyke-rocks from northern Greenland. Data are normalized to average primitive mantle of Sun & McDonough (1989). (b) Extended trace-element normalized diagram for representative sample of mafic dyke-rock from northern Greenland (GGU-434009) compared to an average OIB sample from Sun & McDonough (1989) and the continental crust of Taylor & McLennan (1985). Data are normalized to average primitive mantle of Sun & McDonough (1989).

being either EMI or EMII. Reference to the Ba/Nb and La/Nb plot (Fig. 11b) distinguishes between the latter two sources. The dyke rocks have affinities to an EMI-type source owing to their enrichment in Ba and La (Sun & McDonough 1989), which is in accord with Weaver's (1991) Ba/Th versus Rb/Nb plot (Fig. 11c). Whereas the origin of HIMU basalts is considered to reflect recycling of hydrothermally altered MORB into the mantle via subduction-zone processes, with subsequent genesis of magma after a protracted period of residence, the nature and origin of EM I source regions remain controversial. As recently summarized by Mata *et al.* (1998), the following origins of EM I, which are probably not mutually exclusive, are possible: (1) contamination by pelagic sediments within the subduction-zone environment, (2) delamination or thermal erosion of

subcontinental lithosphere, (3) metasomatism of lithosphere during subduction processes. We do not favor any particular model, but instead emphasize the necessity to have an additional component other than HIMU involved in the source region for the mafic magma. The enrichment of the samples in Ba, La and Rb (Figs. 11b, c) does, however, strongly suggest option 1 above (*i.e.*, contamination by pelagic sediment), whereas the lack of the negative Nb-Ta anomaly would preclude option 3.

The OIB affinity of the dyke rocks is not unexpected, given that the breakup of continents is often accompanied by large amounts of magmatic activity (*e.g.*, Ernst *et al.* 1995). The ensuing rifts commonly form above thermal anomalies, or plumes, in the mantle (White & McKenzie 1989). The generation of these plumes provides a transport mechanism for ascent of OIB-type melts into continental regions. An example of dyke formation similar to that of northernmost Greenland is documented by Storey *et al.* (1999) for the mid-Cretaceous mafic dykes associated with rifting of Antarctica and New Zealand. Of particular interest in their study is the older suite (*ca.* 107 Ma) of mafic dykes, interpreted to have formed from mixing of tholeiitic, OIB and alkaline high- to low-Ti alkaline magmas, which is not unlike the apparently mixed nature of the dyke rocks in this study. In addition, the formation of that dyke swarm was followed some 10 Ma later by peralkaline magmatism (Storey *et al.* 1999), which parallels the subsequent emplacement of the peralkaline Kap Washington volcanic suite in northernmost Greenland.

Of particular relevance to the plume model suggested here are the T_{DM} model ages of *ca.* 720 Ma for the samples, which suggest formation of the source reservoir for the dyke-producing magma during the late Precambrian. This age is coincident with formation of the Franklinian dyke swarm of *ca.* 750 Ma (Trettin 1989), the presence of which is recorded in part along the north-western margin of Greenland (*e.g.*, Thule basin), which is probably a precursor to the continental extension associated with the formation of the Franklinian Basin (Surlyk 1991). Thus, the source reservoir for the dyke-producing magma may have been generated during an earlier extensional event and remained isolated until a later melting event during the Cretaceous, as recorded by the dyke swarm in northernmost Greenland.

SUMMARY AND CONCLUSIONS

A suite of Cretaceous dyke-rock samples from northernmost Greenland that form part of an aurally extensive dyke-swarm have been dated by the $^{40}\text{Ar}/^{39}\text{Ar}$ technique and analyzed for their major- and trace-element chemistry. These data and isotopic (O, Sr, and Nd) abundances suggest the following:

1. The dyke rocks are *ca.* 80–85 Ma, some 20 Ma older than previous whole-rock K–Ar ages suggested. These ages indicate a possible relationship between

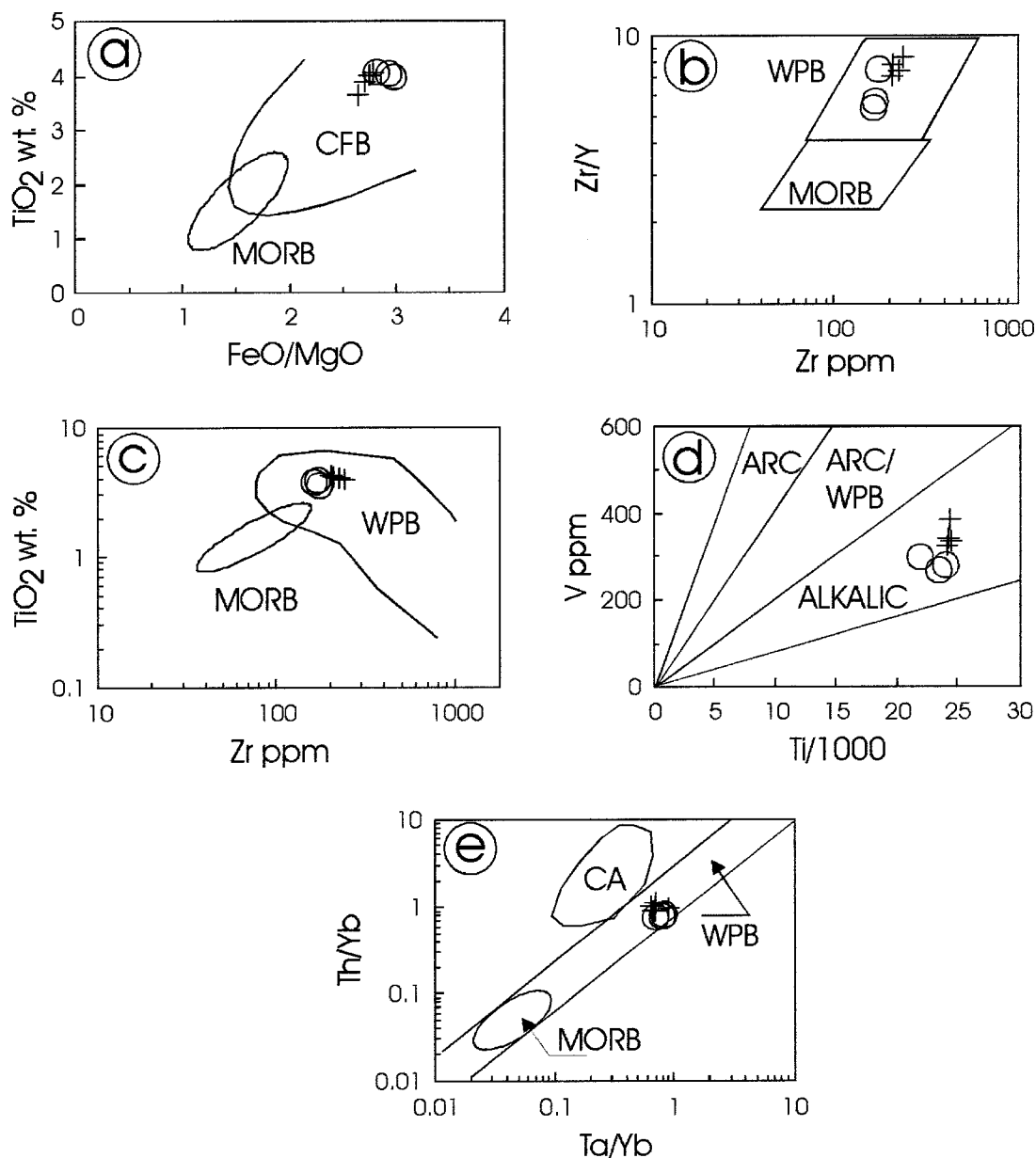


FIG. 14. Binary plots for major- and trace-element data for whole-rock samples of mafic dyke-rocks from northern Greenland in a variety of tectonic discriminant diagrams. Note that sample GGU-434207 has been omitted from the plots owing to alteration. Abbreviations are as follows: CFB: continental flood basalts, MORB: mid-ocean ridge basalts, WPB: within-plate basalts, CA: continental arc. (a) The TiO_2 versus FeO/MgO diagram of Fodor & Vetter (1984). (b) The Zr/Y versus Zr diagram of Pearce (1982). Note that the field for island-arc basalts, which plots to lower values of Zr , has been omitted for simplification. (c) The TiO_2 versus Zr plot of Pearce (1982). Note that the field for arc basalts, which plot at lower values of TiO_2 and Zr , has been omitted for simplification. (d) The V versus Ti/1000 diagram of Shervais (1982). (e) The Th/Yb versus Ta/Yb diagram of Pearce (1983). Note that the field for oceanic arc, which plots in the area occupied by the letters MORB, is omitted for simplification. The position of the Greenland mafic dyke-rocks in this plot suggests a combination of fractional crystallization and within-plate enrichment, as discussed by Pearce (1983).

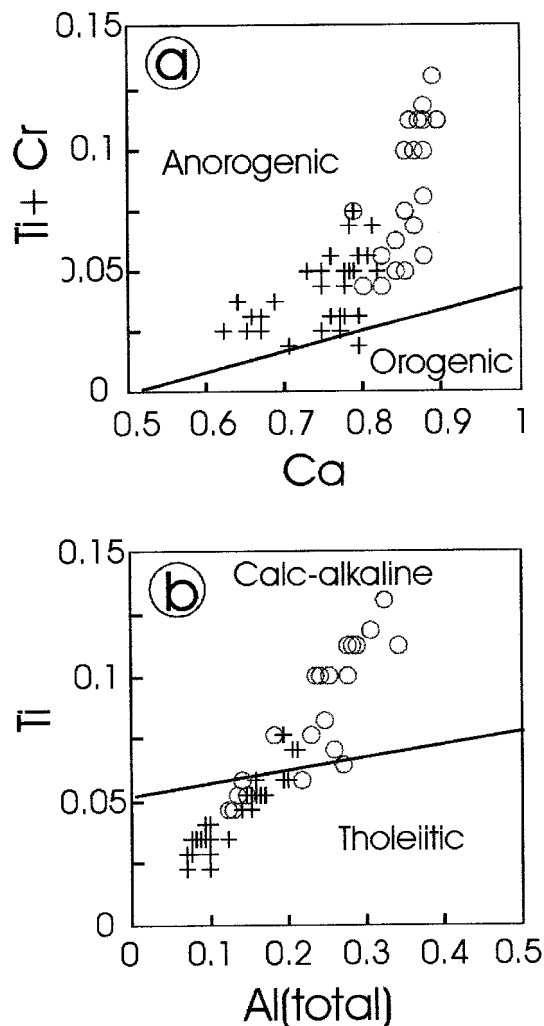


FIG. 15. Tectonic discrimination diagrams (after Leterrier *et al.* 1982), based on the composition of clinopyroxene (calculated on the basis of six atoms of oxygen), for mafic dyke-rocks from northern Greenland. (a) Ti + Cr versus Ca; (b) Ti versus Al(total).

dyke-rock emplacement and ocean-floor spreading off northern Greenland.

2. The dykes rocks apparently have both tholeiitic and alkali basalt affinity on the basis of petrographic and chemical features. They are, however, spatially and temporally affiliated, on the basis of sampling and dating.

3. The dyke rocks have chemical affinities to OIB, were generated at depths ≤ 60 –80 km, and may owe their origin in part to a mantle plume that relates to the development of the Makarov Basin.

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